HOLD PENDING PUB 311
9-23-82

NAGW-133 91

Pergamon '81.
P. 1237-1263

10. Implications of basaltic volcanism for the evolution of planetary bodies

William M. Kaula, Team Leader
—Department of Earth and Space Sciences
University of California, Los Angeles

Fraser P. Fanale
—Institute for Planetary Geoscience
University of Hawaii

Don L. Anderson
—Division of Geological and Planetary Sciences
California Institute of Technology

10.1 INTRODUCTION

The purpose of this chapter is to place basaltic volcanism in the larger context of planetary evolution. We will attempt to synthesize the findings of the preceding chapters into a discussion of the overall evolution of terrestrial bodies. (For a summary of each chapter, see the Overview.) Although our treatment will have this broader point of view, we will avoid excessive bulk by emphasizing those aspects which are related to basaltic volcanism. The plan of this chapter is to describe scenarios of planetary evolution, incorporating other data and ideas with those from the earlier parts of the book, and then to seek heuristic patterns, to infer the leading questions remaining, and to suggest how to pursue them.

Section 10.2 is a discussion of solar system origin as it applies to the terrestrial planets in general. This phase has negligible direct connection to observable consequences of volcanism on the planets, of course. However, consideration of origin is necessary for any comparative discussion of planetary evolution, because there are significant differences in planetary properties which result in different volcanic products, but which must be attributed to circumstances of formation. We identify the boundary between "formation" and "evolution" as the time when the driving energy for major planetary changes shifted from short-lived to long-lived sources. The very early differentiation of the Moon's crust required by isotopic constraints suggests that primordial crustal separation is part of "formation" by the foregoing definition. The circumstances of initial crustal differentiation on all the planets are obscure enough that a comparative discussion is desirable. Another topic appropriately included in this section is basaltic volcanism on small bodies: in particular, the basaltic achondrite parent body and asteroids, not only because the energy sources were probably short-lived, but also because differentiation in small bodies may have a significant effect on the fractionation of protoplanetary material, affecting the bulk composition of the smaller planets (the Moon and Mercury) and the heterogeneity of the larger planets. The compositional consequences of these nebula and planetesimal processes are discussed in section 4.3.

Section 10.3 is a planet-by-planet discussion of subsequent terrestrial planet evolution, which in all cases except Io appears to be internally energized. It is our belief that the considerable differences among the planets and the present levels of understanding of terrestrial bodies are such that it is not fruitful to force a discussion of planetary evolution into a comparative framework *ab initio*. Instead, the main emphasis will be on the thermal, compositional, and tectonic evolution of the planets individually, utilizing the data and ideas presented in the other chapters to the maximum extent feasible.

Section 10.4 is an attempt to draw the leading inferences from the aforedescribed discussion. Although the main theme appears to be that the duration and complexity of a planet's evolution are both positively correlated with its size, there are significant variations from this pattern which may be inherent in planetary behavior rather than just dependent on differences in bulk properties of the planets. We include our opinions as to major problems remaining, and suggest what is needed to progress toward the solution of these problems.

10.2 ORIGIN AND EARLY EVOLUTION OF THE PLANETS

The formation of the planets is generally thought to be associated with formation of the sun, because of the paucity of interstellar material. There is, as yet, no reason to believe that the sun's formation was markedly different from the current formation of similar sized stars within the galaxy. Hypotheses of star formation are constrained to some extent by observations. It is desirable to start from this astronomical base before moving to the speculative matter of planetary formation, and then to end with aspects which are in some ways constrained by planetary, asteroidal and meteoritic properties.

10.2.1 Sun's formation and early evolution

Recently-formed stars are always observed to be close to clouds of gas and dust. This observation, together with the lack of a plausible alternative, leads to the consensus that stars are formed from interstellar clouds. These clouds are probably debris from nova and supernova explosions, because the solar system material contains heavy elements which require processes within giant stars for their synthesis (Trimble, 1975). The duration and cessation of nucleosynthesis and the number of sources affect the products of short-lived radioactive decays: ^{26}Mg, ^{129}Xe, ^{134}Xe, ^{136}Xe, and others recently identified. The **formation interval**, or the time from the end of nucleosynthesis to entrapment of the daughter isotopes in meteoritic material, is about 180 m.y. from xenon isotopes produced by ^{129}I and ^{244}Pu decay (Schramm *et al.*, 1970). However, the detection of ^{26}Mg excesses in the Allende meteorite, indicating the presence of the short-lived radioisotope ^{26}Al (Lee *et al.*, 1976; Gray and Compston, 1976), has raised the suggestion of a second source with a much shorter formation interval of 2 or 3 m.y. (Cameron and Truran, 1977). Another type of constraint is furnished by varying ratios in the stable isotopes, most notably $^{18}O/^{16}O$ and $^{17}O/^{16}O$ (Clayton and Mayeda, 1975; Clayton *et al.*, 1976). As discussed in section 7.2.2, a plot of variations within any one planetary body shows a distinct 2:1 trend of the two isotope ratios, explicable by diffusive and chemical mass fractionation effects. However, there are other differences among bodies explicable only by preservation of the consequences of incomplete mixing of materials from different nucleosynthetic sources. Isotopic differences among bodies (other than dispersion along the 2:1 line) then require that these bodies come from different parts of the solar system raw material. Conversely, agreement of isotopic ratios of two bodies suggests they came from the same part (but does not require it). Lunar rocks fall along the same 2:1 line as terrestrial rocks, and have a centroid close to that of terrestrial igneous rocks. Basaltic achondrites fall on a nearby parallel line but have a distinctly different centroid. Ordinary and carbonaceous chondritic meteorites all fall appreciably farther from the terrestrial line.

For a star to form from a gas-and-dust cloud, effects inducing contraction (gravitation, external pressure) must overcome those maintaining extension (temperature, angular momentum, turbulence, magnetic field). Considering only gravitation and temperature, a spherically symmetric isothermal cloud in equilibrium must have a density varying with radius, $\rho \propto r^{-2}$, to create a pressure gradient counterbalancing gravitational attraction. Under these extremely idealized conditions, the requirement for collapse is

$$M > 2c^2R/G \tag{10.2.1}$$

where M is mass, R is the radius of the whole cloud, G is the gravitational constant, and c is the sound speed, proportionate to $T^{1/2}$. Some clouds are observed to have a temperature T as low as 10° K; the condition (10.2.1) for a $1 M_\odot$ mass then requires a mean density ρ greater than 10^{-19} g/cm^3 ($M_\odot$ is a solar mass). While some small dark clouds have such density and temperature, it is not known whether they can collapse, or whether stars as small as the sun must form as fragments of a much larger cloud, in which external pressures work with gravitation to collapse smaller fragments. The formation of a smallish star like the sun probably entails appreciable asymmetry of conditions, and may normally occur only within a cluster of stars.

The computed collapse of an isolated spherically symmetric star from 10^{-19} g/cm^3 takes ~1 m.y., even without considering rotation or magnetic fields. Hence a significantly more rapid collapse requires an extraordinary external pressure.

Forming stars are identifiable because they are cooler, and hence redder, than normal stars of the same brightness. Spectral observations of forming stars indicate appreciable mass motions and heating in their vicinity. Some of the motions are interpreted as being mass outflows from the stars, in the form of T Tauri winds, on the order of 10^7 times the present solar wind, which is $\sim 10^{-14} M_\odot$/year. The heating is often interpreted as the consequence of material infalling to an **accretion disk**: an extended sheet of matter around the star. All these matters are the subject of appreciable research and debate (Strom, 1977; Ulrich, 1978).

The sun probably separated from its original cluster of stars and settled down to its long term behavior within 1 m.y. or so. Although it has undergone significant evolution (spin down due to loss of angular momentum with solar wind, heating due to increase of mean nuclear mass with hydrogen burning), the sun probably was not a major agent for moving matter or internal heating of solid bodies after this initial epoch.

In summary, there is appreciable rationale for external effects on planetary material, both from the forming sun and from outside the solar system, in the first ~30 m.y. Thereafter, conditions were essentially similar to current conditions: an isolated solar system with a sun acting mainly as a source of gravitational attraction and radiation in the visible part of the electromagnetic spectrum.

10.2.2 Solar nebular models; gravitational instability

Given the existence of a rotating accretion disk with some means of internal transfer of momentum, such as turbulent viscosity or hydromagnetic coupling, then there inevitably must be a net inward transfer of mass and a net outward transfer of angular momentum. The torque between adjacent material in the nebula must go to zero with the rotation gradient $d\Omega/dr$ at the co-rotating sun, while it must go to zero at the outer limits because the coupling effect (whether mechanical or magnetic) goes to zero (Lynden-Bell and Pringle, 1974). At an intermediate distance, there must be a maximum torque; to compensate the resulting outward flow of high angular momentum matter, conservation requires an inward flow of low angular momentum matter. Other likely properties of such an accretion disk are heating from conversion of the kinetic energy of infalling material, with consequent loss of gas, and, if of sufficient density, the formation of mass agglomerations by gravitational instability. If the gaseous disk is not sufficiently dense, these agglomerations may arise in a layer of condensed matter.

The protoplanetary material, commonly known as the **solar nebula**, plausibly was an accretion disk. The greatest uncertainties about fundamental determiners of this accretion disk are (1) the rate of infall of material into the protosolar system; (2) the relative phasing of formation of the sun and the occurrence of major gravitational instabilities in the nebula; (3) the amount and timing of "T Tauri" mass outflows from the early sun. "Minimal" theories, such as Safronov's (1972), which hypothesize a nebula a few times as massive as the planets plus their volatile solar complement ($\sim 0.015 M_\odot$), entail a slow rate of infall comparable to that calculated for formation of an isolated $1 M_\odot$ star, formation of the bulk of the sun before any significant agglomeration in the nebula, and slight outflow of mass from the sun. "Maximal" theories, such as Cameron's (1978a), which hypothesize a $\sim 2 M_\odot$ nebula, entail an infall rate ~ 30 times as great, formation of major gaseous gravitational instabilities before the existence of a large solar core, and a great subsequent mass loss from both solar outflow and accretional heating.

The gravitational instability equation (10.2.1), can be converted to a relation between a characteristic length, $\lambda \sim R$, and density, $\sim M/R^3$. For application in the solar nebula with angular rate Ω at radial distance r, two further modifications are desirable: (1) The flattened shape of the nebula makes a surface density $\sigma \sim \rho H$ appropriate, where H is a scale height normal to the plane of the nebula: $H \sim c/\Omega$. (2) The rotation of the nebula necessitates taking into account the shearing effect of a gradient in angular momentum, $d(\Omega r^2)/dr$. A consequence of this latter effect is to set an upper and a lower bound on the size of an unstable agglomeration (Safronov, 1972; Goldreich and Ward, 1973)

$$\frac{2\pi^2 G\sigma}{2\Omega^2 + r\Omega d\Omega/dr} > \lambda > \frac{c^2}{G\sigma} \qquad (10.2.2)$$

For gas, the velocity c is the thermal velocity, as used in Eq. (10.2.1). For solids, however, the velocity c is the turbulent velocity. Whether the instability which occurs is of the gas or the dust hence depends on temperature and turbulence as well as surface density σ. The minimal theories attain densities σ sufficient only for dust instability, while the maximal theories could attain gas instability. The minimal theories are more easily reconcilable with meteorite and terrestrial planet evidence. However, maximal theories explain agglomeration of the H + He rich planets, Jupiter and Saturn, more satisfactorily. Therefore, it is desirable to examine the implications and variations of both hypotheses for bodies on which basaltic volcanism has occurred.

10.2.3 Dust instability and planetesimal growth hypothesis

In this scenario, as outlined in greatest detail by Safronov (1972; see also Harris, 1978), dust grains formed by condensation settle to the central plane of the nebula in a time on the order of 1000 years. The initial planetesimals for a minimal nebula are on the order of 1 km size at 1AU distance; if gas drag or other dissipative mechanism is operating, these bodies form aggregations of ~ 10 km. This process is inexorable, so long as the time scale of reduction of mutual velocities by bumping and gas drag is shorter than the time scale of disruption due to infall of matter into the nebula or to solar activity. The relative velocities in gravitational instability are small—about 10 cm/sec. Hence the initial agglomerations were probably rather loosely put together and of low density.

Subsequent to formation of the initial planetesimals by gravitational instability, there is a growth of both planetesimal masses, due to collisions, and relative velocities, due to mutual perturbations. The rate of this growth and the velocity and mass distributions at various stages are somewhat uncertain, because the principal analytic treatment by Safronov (1972) separates mass growth from velocity growth, while computer experiments to date have been limited to extreme circumstances: uniform-sized starting populations of planetesimals, and either only the very early stages (Greenberg *et al.*, 1978) or only the very late stages (L. P. Cox, pers. comm., 1980; Wetherill, 1980a).

In any case, some plausible constraints can be stated: (1) The time scale of growth of planets apparently was long compared to the time scale of removal of volatiles from the zones of formation of proto-Earth planetesimals and meteorite parent bodies. Otherwise, the amount of volatiles swept up and held gravitationally would have been significantly greater. This constraint is most severe for Xe, for which at 500°K a planetesimal of ~ 100 km radius in a medium of solar composition, Xe:Si $\sim 10^{-6}$, would acquire more than a terrestrial or meteoritic Xe:Si ratio, $\sim 10^{-11}$. (2) The material of chondrites, once collected, cannot have been subjected to high velocity ($\gtrsim$1km/sec) impacts. Otherwise, they would evidence appreciable shock effects. (3) Either the temperature remained above $\sim$1360° K in the zone of proto-Mercury material until volatiles were removed, or subsequently some sort of mechanism acted to remove silicate-rich material in preference to iron-rich. Otherwise, the planet Mercury would have a significantly lower density. (4) Relative velocities in the later stages of planet growth were at least as great as the average noncircular components of the orbits of the Earth and Venus themselves, $\sim$1km/sec. (5) Jupiter probably had a much faster growth rate than the terrestrial planets, to account for the acquisition of H + He by Jupiter before the T Tauri wind blew it away, the absence of a planet in the asteroidal zone, the high relative velocities of asteroids, and the stunting of the growth of Mars. The dearth of matter and the amount of random motion in the asteroid-Mars zones both suggest some strong dynamical disturbance, such as might have been generated by a forming Jupiter (Weidenschilling, 1975).

Models of the evolution of a planetesimal population generally obtain relative velocities moderately smaller than the escape velocities of the average mass $\overline{m}$ of the population, $\sim(2G\overline{m}/\overline{r})^{1/2}$. The resulting capture cross sections for bodies approaching the terrestrial planets in size led to growth times well over 1 m.y. To inhibit the growth of a sizeable planet in Mercury's zone there must be either some sort of disturbing effects from the sun (Safronov, 1972) or a mechanism to remove a large portion of the planetesimals, such as a gas drag which would act more strongly on bodies of large area: mass ratio; i.e., small and silicate-rich, rather than massive and iron-rich, bodies (Weidenschilling, 1978).

Since it is a satellite, the Moon requires some special explanation. The extreme iron and refractory siderophile deficiency of the Moon (section 4.5.2) suggests previous planetary differentiation. Dynamically plausible mechanisms which minimize the energy requirements to place protolunar material in geocentric orbit indicate that this planetary differentiation took place in the Earth or in bodies in the Earth's vicinity. Oxygen isotope ratios measured in lunar and terrestrial rocks are very similar to one another, but differ markedly from those seen in meteorites, which lends support to the dynamical arguments. It is unlikely that the differentiation took place in protomoons in geocentric orbits, since known mechanisms (tidal friction, gas drag, collisions) would all act to remove the lighter, silicate-rich objects and retain the heavier, iron-rich objects in orbit around the Earth. There remains the Earth or other bodies in heliocentric orbits. Both loci involve dynamical difficulties in placing iron-deficient material in geocentric orbits. A strong argument against the differentiation having occurred in other bodies in heliocentric orbits is the absence of meteorites which appear to come from the same compositional pool as the Moon (Kaula, 1977; Ringwood, 1979; Drake, 1980).

Removal of the lunar material from the early Earth has the obvious attraction of utilizing a well established condition: a separate core and mantle at some early stage. The composition of the Moon is thus accounted for by removal of the protolunar material from the mantle after core separation. This removal may have happened well before the completion of the Earth's growth. Two mechanisms have been proposed for the removal: (1) rotational instability (O'Keefe and Sullivan, 1978); and (2) collision by another planet-sized body (Hartman and Davis, 1975; Cameron and Ward, 1976). Objections to the rotational instability hypothesis are (a) to provide sufficient energy, core formation would have to be delayed until the Earth was almost entirely formed, which in turn requires keeping the forming Earth unreasonably cool (see below); and (b) about ten lunar masses and an even greater proportion of angular momentum must be ejected from the Earth-Moon system, since the fundamental mode of rotational instability is the lowest degree non-axisymmetric distribution; i.e., a third harmonic. The main objection to the collision hypothesis is the seeming *ad hoc* character of requiring such a great impact. The terrestrial bodies are a disparate group of n = 5 bodies, however, and it should not be at all surprising if there were once six, seven, or more bodies of size comparable to Mars or Mercury (Wetherill, 1976). An additional argument in favor of the collision hypothesis is that the material is more likely to be placed in a dispersed state, so that volatiles can be lost (Kaula, 1977; Ringwood, 1979).

Acquisition of protolunar material from collisions of bodies in heliocentric orbit does not entail any severe intrinsic difficulty. Mechanisms for reducing the iron content at capture, or subsequently by additional collisions, however, are rather speculative (Ruskol, 1977; Kaula and Harris, 1975).

The minimal nebula hypothesis is a congenial working model for discussing variations in bulk composition (see Chapter 4), phasing of early phenomena (see Chapter 7), and initial thermal state (see Chapter 9) among the terrestrial bodies: so congenial it is often assumed implicitly. The hypothesis, however, does have at least two major problems: (1) How did the Jupiter and Saturn zones develop cores large enough to capture a large amount of hydrogen and helium before it was swept out? The obvious suggestion is that these zones very early had a temperature low enough for ice to condense, both H_2O and CO_2. But this turns out to be a rather marginal advantage over the terrestrial zones for any plausible density gradient (i.e., negative $d\sigma/dr$) in the nebula. (2) Why is there a steep dropoff in inert gas content in the sequence Venus-Earth-Mars? In primordial argon, $^{38}Ar + ^{36}Ar$, the proportionate abundance ratios are 20,000:200:1. These ratios are much too great for some sort of pressure gradient in an isothermal nebula (Pollack and Black, 1979). Alternatives are (1) a steep radial gradient in solar wind implantation or (2) impact by a large body, or bodies, from closer to the sun (Wetherill, 1980b). Since the implantation gradient hypothesis requires appreciable inhibition of subsequent dynamical mixing, the large impacter seems more plausible.

10.2.4 Gas instability and giant protoplanet hypothesis

In this scenario, as described by Cameron (1978a, b), the density of the nebula becomes so great that instability of the gas can occur. For Jupiter at $\sim 140°$ K, forming appreciably further out in the nebula than its present location, the required surface density σ is ~ 500 g-cm^{-2}. The present Jupiter with a ~ 15 earthmass icy and rocky core plus its solar complement spread from ~ 10 to 24 AU would be ~ 35 g-cm^{-2}. After stage (1), collapse of gas-plus-dust, evolution to the present major planets requires the following stages: (2) compositional segregation within the agglomeration, to obtain an ice- and rock-rich core; (3) stripping off most of the gaseous outer parts; and (4) removal of this excess gas from the solar nebula. Stage (2) would occur as the consequence of condensation and "rainout" within the agglomeration (Slattery, 1978; Slattery *et al.*, 1980), somewhat similar in physics to rain in the Earth's atmosphere. Stage (3) is the greatest difficulty. It is hypothesed by Cameron (1978b) to arise from a later growth of the solar core than the protoplanetary agglomeration, with a slower collapse time of the latter, so that the protosun can tidally disrupt the protoplanet. Stage (4) can be the result of the evolution of the accretion disk, in which part of the mass spirals into the sun, part is lost by "evaporation" from the outer parts, and part may be swept out by a protosolar wind.

Although the motivation for the giant protoplanet hypothesis comes from the outer solar system and astronomical considerations, its implications for the terrestrial bodies should be examined. At least three distinct scenarios can be conjectured. (1) The terrestrial planets underwent the same four stages as Jupiter and Saturn, differing mainly in having a much more drastic stripping off of their gaseous envelopes. (2) The terrestrial planets underwent stages (1) and (2), gas instability and compositional segregation, but then in addition to stripping off of gaseous envelopes, the solid cores were broken up to form planetesimals, which subsequently recombined to form planets. (3) The terrestrial planets never underwent gaseous instability, but only dust instability. In these scenarios, the planets formed considerably further from the sun than their present locations, because to lose mass from the nebula, there must be a proportionately greater loss of energy and angular momentum.

All of these scenarios require modifications to explain how Mercury, Mars, and the asteroid belt are different from the Earth and Venus. A major difficulty with scenario (1) might be the drastically greater depletions of the terrestrial bodies in inert gases than in active gases. While any segregation process would carry down with the solids more active gases than inert gases, it is difficult to imagine it being so efficiently discriminatory. Another difficulty of scenario (1) is that a mighty blast is required to remove virtually all the gaseous envelopes; however, such a blast might also account for the mercurian metal:silicate ratio. Scenario (2) is justified to explain the great depletions of all volatiles, but its energetic requirements for breaking up cores are even more extreme than those of scenario (1). Scenario (3) entails appreciably greater turbulence in the inner solar system or some sort of turbulence or accretion disk torque effect to inhibit gaseous instability, but none of these are wildly implausible. All three scenarios are difficult to reconcile with the fragile, low temperature properties of carbonaceous chondrite meteorites, which look much more like primordial than recondensed objects. In a hypothesis which requires late formation of the sun, however, it is a minor wrinkle to bring meteoritic material into the solar system at a later stage when the activity has calmed down.

The compatibility of gaseous instability hypotheses with terrestrial body properties is as yet little explored. Until a better explanation is found for the much larger scale phenomena of Jupiter and Saturn formation, however, this exploration should be continued.

10.2.5 Small body heating and differentiation

A feature of the processes leading to planetary formation is the heating of planetesimals and their resultant differentiation. Small body differentiation has produced basaltic volcanism in at least two instances: the eucrite parent body and the asteroid Vesta (which may be the same body: Consolmagno and Drake, 1977). Small body differentiation also is of interest as being an important potential contributor to differences in planetary bulk composition through collisional differentiation, as discussed in section 4.3.2; this mechanism is particularly important for any nonfission origin of the Moon. A problem for any sort of small body heating is the energy source. A source which necessarily existed is collisions. To convert a significant amount of relative kinetic energy into heat, however, requires impact velocities of well over a km/sec (O'Keefe and Ahrens, 1977). So long as planetesimal growth had not progressed beyond the creation of bodies with escape velocities of more than a km/sec (i.e., more than ~750 km in diameter), such velocities must have been rare indeed; otherwise, the planets never would have collected. While it is not to be ruled out that occasional higher impact velocities occurred as the result of injecting bodies from outside the immediate circum-solar zone (Kaula and Bigeleisen, 1975; Weidenschilling, 1975), the small portion of shocked meteorites and the exceptionality of Vesta's spectrum indicate that hypervelocity impacts between smallish planetesimals were unusual.

It has been an objection to collisional heating that the textures of the melted material would be much more brecciated and glassy than those of basaltic achondrite meteorites or other basalts. It also is difficult to imagine the classic processes of magmatic fractionation being mimicked in the few milliseconds of the high pressure phase of an impact. Experience with the lunar rocks, however, has led to the identification of more and more differentiated specimens as impact melts, so that even KREEP basalts have been hypothesized to be greatly modified by impact melting (Warren and Wasson, 1979). Collisional effects may also be important for compositional differentiation among meteorites, which in turn has implications for planetary composition and evolution, as discussed in sections 4.3.2 and 10.2.7.

The above doubts about the effectiveness of collisions have led to two other hypotheses of planetary heating, electromagnetic induction and radioactive decay. Electromagnetic induction arises from the intersection of magnetic field lines by an electrically conducting body (Sonett and Herbert, 1977). The difficulties of the hypothesis are twofold: firstly generating a sufficiently intense and rapidly changing magnetic field, and secondly raising the temperature of the planetesimal sufficiently (a few 100° C) by some other mechanism to give it an appreciable electrical conductivity. The intensity of the early solar magnetic field indicated by the remanent magnetism of meteorites, 1 gauss (Butler, 1972), may be adequate. More conjectural is whether there was a sufficiently rapid alternation of the magnetization, most plausibly arising from a sectored structure of the solar magnetic field and a rapid solar spin.

The one outstanding candidate for radiometric heating sufficient to melt planetesimals within the first few million years is ^{26}Al decay, which has a half life of 0.75 m.y. Originally proposed by Urey (1956) and elaborated by Fish *et al.* (1960), the hypothesis has been revived by discovery of a ^{26}Mg excess in inclusions of the carbonaceous chondrite Allende (Lee *et al.*, 1976; Gray and Compston, 1976). From this limited data on a few meteorite samples, it is difficult to quantify the probable significance of ^{26}Al as an early heat source for small bodies. As mentioned in section 10.2.1, the anomaly is one of those which indicate compositional heterogeneity in the solar nebula.

The intrinsic properties of eucrites, as discussed in sections 1.2.8 and 7.2, afford some appreciable constraints on the nature of their parent body. The source region of eucrites was quite dry and had metal present (Stolper, 1977; see also section 3.5). The bulk composition of the parent body could plausibly have been chondritic (Consolmagno and Drake, 1977). The spectacular depletions in refractory siderophiles (Morgan *et al.*, 1978) indicate a rather slow acting process, such as would result from a small percentage partial melt at depth, rather than a "lake" of impact melt.

The distinctly basaltic spectral signature of Vesta, discussed in section 2.2.6, may result from exceptional circumstances. Somehow, most of the original material in the asteroidal zone was cleared out. Since the escape velocity from the asteroidal belt is at least 4 km/sec, it is surprising that only a small minority of asteroids show differentiated surfaces, with Vesta's the sole basalt-like spectrum (section 2.2.6; Chapman, 1979). The differentiation of Vesta suggests either collisional heating (since Vesta is rather remote from the sun for electromagnetic induction), or a sufficiently earlier formation of Vesta compared to most other asteroids to yield appreciably greater ^{26}Al heating.

10.2.6 Initial energetics of terrestrial planets

The temperature of formation of a planet (to be distinguished from the temperature of condensation of its material) depends on, in probable order of significance: *(a)* the size of the planet, *(b)* the sizes of the bodies from which it is formed, *(c)* the heliocentric velocities of the forming bodies, *(d)* the composition of the planet, *(e)* its heterogeneity, *(f)* its epoch of formation, *(g)* its distance from the sun, and *(h)* its rate of formation.

(a) The Mass M of the planet obviously acts by its influence on the impact velocities of infalling matter, the kinetic energy per unit mass being largely GM(r)/r. Integrating over r from 0 to R gives as the mean energy/unit mass of a homogeneous body 3GM(R)/5R. This zero order result suggests that the initial mean temperatures of planets will vary as $M^{2/3}$, and that the temperature within a given planet will vary as r^2, so long as there is no convection or differentiation. The mean energy release per unit mass upon core formation would also be expected to vary proportionate to gR, or $M^{2/3}$. Core formation would reverse the temperature profile in the mantle, since the further a material element travelled in the differentiation process, the more energy it would have gained on the average, because of greater interaction with its surroundings. This consideration applies to upwardly moving, as well as downwardly moving, elements. Core separation probably got underway before the completion of planet formation for any body larger than the Moon.

(b) The principal neglect in the foregoing estimate is the loss of heat by radiation at the surface, $\sigma(T_s^4 - T_0^4)$ in energy/area/time. The strong dependence on surface temperature T_s means that normally the heat loss is governed by the rate of heat transfer from the interior to the surface, -KdT/dr. The main variable affecting the gradient dT/dr is the size of the infalling bodies: the larger the bodies, the deeper the deposition of energy. Infalling bodies also act to transfer heat outward by stirring up the planetary material. However, this effect is limited by the relatively small partitioning in hypervelocity impacts of the kinetic energy of the impacter to kinetic energy of ejecta, and attenuates more rapidly with respect to depth than heat deposition. In general, therefore, there should be a positive correlation of heating with size of body forming the planet. A difficulty in quantifying this result is the portion of energy lost by radiation from fragments created at the time of impact. The retained h energy fraction necessary to attain a prescribed temperature T at a given radius r can be obtained from, roughly,

$$h\frac{GM(r)}{r} = \rho C_p[T(r) - T_0] \qquad (10.2.3)$$

The partitioning factor h for retained heat should have a positive correlation with size of impacts, probably attaining more than 0.5 in large events (Kaula, 1980a).

As discussed in sections 10.2.2–3 above, the size of bodies forming planets depends strongly on whether the initial instability was dusty or gaseous.

(c) The heliocentric velocities of the forming bodies could have significantly contributed to the impact energy, if relative velocities were on the order of kilometers/second.

(d) The composition of a planet affects its initial heating mainly through its effect on core formation, which requires the presence of metallic iron or iron sulphide. This presence requires either a deficiency of oxygen at the time the temperature drops low enough for iron oxides to be stable (~600° K), or the introduction of reducing agent, such as carbonaceous material. In the case of Mercury, there must also be a deficiency of silica (SiO_2) when the temperature drops below the condensation point of fayalite, ~1360° K. The most marked evidence of the thermal effect of planetary composition is the different evolutions of Mercury and Mars. The initial hot state, low oxidation level, and high iron content of Mercury's material led to rapid core formation and a predominantly cooling history. The more carbonaceous-chondrite-like composition of Mars apparently led to late core formation and a mainly tensional tectonic regime (sections 9.5.3–4).

(e) The formation temperature of a planet will vary inversely with its degree of heterogeneity. A declining iron portion in time of the accreting body will accordingly decrease the amount of gravitational energy available from core formation. Any lateral heterogeneity, such as large planetesimal infall or instabilities associated with core formation, will induce convection and volcanism, leading to more rapid heat loss.

(f) The epoch of formation of the planet is critical not only because of solar-dependent processes, such as the ambient temperature T_0 and the removal of volatiles alluded to in section 10.3.4 above, but also because of the possible presence of short-lived radionuclides, such as ^{26}Al, discussed in section 10.2.5.

(g) The distance from the sun has both thermal and dynamical effects. The thermal effects are manifested mainly in planetary compositions, as discussed in section 4.5. Only in the case of Mercury might solar insolation have a significant effect on a planetary thermal state. Probably more important were dynamical effects, since orbital velocities vary inversely with solar distance. In addition, there were interaction of a strong solar wind with the nebula (Elmegreen, 1978), and drag effects of the nebula gas (Weidenschilling, 1978).

(h) The rate of formation of a planet probably was of comparatively small import, since, as discussed above, the mass was probably acquired mainly as large

planetesimals which bury their heat. In fact, there may be some reason to expect a mild inverse correlation of heating with rate of formation, since a slower rate would occur with larger planetesimals, inhibiting growth due to the enhanced relative velocities (Kaula, 1979). The body most likely to have an initial thermal state enhanced by rapid growth is the Moon, if it were made from debris around the Earth resulting from a major collision.

10.2.7 Early planet differentiation

As discussed in section 4.5, all the terrestrial bodies have significant indication of crustal separation, while three, the Earth, Mars, and Mercury, appear to require a core from either moment of inertia or magnetic field. However, any planet which has been hot enough to differentiate a crust, has a plausible content of radioactive heat sources, and has reduced iron (or sufficient reducing agent present) will differentiate a core. The Moon is anomalous in having very little iron. The Moon has not enough iron in its reduced state to give evidence of a metallic core other than some puzzling remanent magnetism. Also affecting core formation is compositional heterogeneity of infall. The slow cooling rates inferred for iron meteorites (Wood, 1964; Goldstein and Short, 1967) indicate that there existed large heterogeneous planetesimals. An unknown of terrestrial planet formation is the degree of heterogeneity: in particular, the sizes of predominantly iron planetesimals within a mainly silicate population. However, the temperature dependence of viscosity, together with the large stresses induced in any silicate center of a protoplanet, appear sufficient for rapid early core formation starting from only 10 km iron concentrations (Stevenson, 1980).

Only for the Earth, Moon, and the eucrite parent body is there radiometric evidence that crustal differentiation occurred very early, and even in the case of the Earth it is an indirect inference (though rather inescapable), as discussed in section 7.4. But these constitute a sufficient range of conditions that it seems plausible to assume that all terrestrial bodies differentiated crusts early. For the Earth and Moon there is also evidence that the amount of crust created early is comparable to the present bulk. Again, this evidence is more clear cut for the Moon, where the terrae lead isotopes point rather conclusively to a general differentiation 4.4 b.y. ago (see section 7.3). Furthermore, given that as small a body as the Moon differentiated a thick crust more than ~ 4.0 b.y. ago, it is thermomechanically most plausible that the energy for the differentiation was delivered within the first ~0.1 b.y. after solar system origin. Hence the energy source could not have been any of the long-lived radioactivities (U, Th, and K) but must have been one of the short-lived processes (collisional, electromagnetic, or radioactive: see sections 10.2.5–6.)

Whether these early crustal differentiations had some similarities to basaltic volcanism as discussed in this book is problematical. The eucritic meteorites evidently came from a body (or bodies) not only appreciably smaller than the planets, but also drier than the Earth and probably more reduced than either the Moon or the outer parts of the Earth. The short time scale of planetary differentiation indicates that it was a much more violent process than the subsequent tectonics and volcanism indicated by the crater count stratigraphy in Fig. 8.9.1. The ~70 km thickness of the lunar crust (or other compositional variation accounting for isostasy) indicates that the differentiation was a global phenomenon, involving a major part of the planet. On the other hand, the ~14 km average thickness of the Earth's crust and the rates of crustal formation in accessible geological time, as discussed in section 6.2, indicate that the Earth has recycled an appreciable amount of crust to the interior. Even a recycling of radioisotopes (Sm/Nd, Rb/Sr) in the first ~2.0 b.y. do not conflict with any of the isotopic indicators of long-term separation of oceanic and continental source regions discussed in section 7.4. The indication of inverse correlation of crustal thickness with planetary size, Table 4.2.1, may be related to either the compressibility of melts (E. Stolper, pers. comm., 1980) or to the basalt-eclogite phase transition and the floating or sinking of anorthite-rich "rockbergs" in a magma ocean. The crusts of both Earth and Moon are about one-fourth as thick as the depth at which the basalt-eclogite phase transition intersects the solidus (Fig. 3.3.27), and hence the depth at which sinking, rather than rising, of plagioclase would occur. The same could be true of Mars, but Venus appears to have a greater ratio of crustal thickness to phase transition depth.

A question closely interacting with early planetary differentiation is that of the degree of compositional heterogeneity in formation. Heterogeneity is most strongly indicated for the Earth, in which a reduced iron core and hydrous outer parts co-exist. However, heterogeneity cannot be ruled out for the other terrestrial bodies. In the dust instability and planetesimal growth hypotheses discussed in section 10.2.3 heterogeneities would arise from cooling of the nebula, differential settling of condensate grains in the nebula, and interzone mixing, such as a greater admixture of volatile material from further out in the solar system in the later stages of growth, as suggested by Anders and Owen

(1977). Dynamical rationales can be contrived to explain the dearth of active volatiles in Mars and the Moon, but not for the aforementioned drastic differences in primordial inert gas abundances. In the gas instability and giant protoplanet hypotheses, heterogeneity would be a plausible consequence of the "rainout" of the solid planet from its gaseous envelope due to differences in condensation temperatures, droplet densities, and other factors (Slattery, 1978; Slattery *et al.*, 1980). Such processes could lead to an iron-enriched center and volatile-enriched outer parts, since the rainout rate likely was more rapid than the cooling rate (in part due to the energy resulting from the rainout itself). Where calcaluminous and ferromagnesian silicates would become located relative to each other is more conjectural. Why Mars is small and lacking in volatiles is not clear in this hypothesis, and the Moon must be formed by a separate mechanism, such as a major collision with the Earth.

The relatively small amount of evidence available suggests that the planets were as different in the characters of their initial differentiations as they have been in subsequent compositional evolution. We do not know whether the very early differentiations of the Earth produced material like Archean basalts: 0.8 b.y. is too large an extrapolation range. But the greater heating and volatile content of the Earth suggest that, if the Moon had a "magma ocean," the Earth had an outer layer which was highly disrupted for a much longer time, with appreciable ferromagnesian content and a great deal of recycling: quite different from what the Moon settled down to within 0.1 or 0.2 b.y., if the collected samples of highland rocks are at all representative.

In conclusion, about all that can be suggested is: (1) the outer parts of the terrestrial planets were quite hot in their first few 0.1 b.y., (2) the degree of heating had a strong positive dependence on planetary size, (3) the ability to stabilize an initial differentiated crust plausibly had an inverse correlation with size, and (4) there could have been an initial gradient in composition from more refractory at the center to more volatile in the outer parts in each planet. More uncertain is whether volatile correlation with solar distance or planetary size dominated. Certainly the correlation is predominantly with size out to Mars, and the resulting volatile contents probably had significant influence on crustal stabilization.

10.3 SUBSEQUENT EVOLUTION OF THE PLANETS

As mentioned in the introduction, for the stages of planetary evolution subsequent to the time when short-lived radioactivities, material infalls, and solar processes (such as electromagnetic induction) could have had important influences our knowledge of the differences in planetary character makes separate discussion of each body appropriate.

10.3.1 The Earth

Long after the early heating and differentiation discussed in sections 10.2.6–7, this largest of the terrestrial planets is still convecting vigorously. As reviewed in section 9.5.1, this mantle convection and the resulting plate tectonics are plausibly the consequence of heat which is to a minor (but not to a negligible) fraction primordial and to a major fraction radioactive, aided by a temperature dependent rheology and a thermal expansivity enhanced by phase transitions. The major current debate is over the extent of material exchange between the upper and the lower mantles.

Despite the ample material brought to the surface by this vigorous convection, the present basaltic crust is but a thin veneer. Currently, the Earth is subducting oceanic crust about as fast as it is being formed. If this return is at a rate proportionate to the energy provided to convection, then there must be a large amount of basaltic material in the mantle, as has been pointed out by Dickinson and Luth (1971), Ringwood (1975), and others.

While plate tectonics requires this appreciable circulation of matter, isotopic data indicate that there are several reservoirs in the mantle which have remained separate for the order of 1 to 2 b.y. (or have equivalently slow mixing rates). One of these reservoirs provides the mid-ocean ridge basalts and the other provides the continental flood basalts, and possibly one component of the blend that supplies the oceanic islands.

The mantle is heterogeneous in chemistry, isotopic composition and seismic velocities. But it is not clear whether or how the seismological and geochemical heterogeneities are related. The largest lateral variations in seismic velocities occur in the outer 200–250 km of the Earth and are related to such surface tectonic features as shields, trenches, rises, and volcanic belts, as discussed in section 6.2. The mantle also has radially inhomogeneous rheology, with the lithosphere, asthenosphere, and transition zone being the main subdivisions of the upper mantle. It also is not clear whether the chemical differences, particularly evident in the trace element and isotope geochemistry, are related to the lateral or radial variations in mantle structure.

Ocean floor basalts have comparatively uniform and low large-ion radioisotope content: $^{87}Sr/^{86}Sr < 0.703$, $^{206}Pb/^{204}Pb < 18.7$, and $^{144}Nd/^{143}Nd > 0.51305$, whereas continental basalts and basalts from ocean islands not associated with island arcs have less uniform and higher ratios (see section 7.4.2). In some cases there is close geographic proximity of lavas derived from distinctly different source regions. Radioisotopes impose a time constraint. Reservoirs with different element ratios—Rb/Sr, U/Pb, Th/Pb, and Sm/Nd have existed for the order of 1 to 2 b.y.—ten times older than present oceanic plates. The source region for mid-ocean ridge basalts (MORB) provides rather uniform composition lavas throughout the world. It must therefore be immense in size and global in nature. On the other hand, continental and ocean island basalts, although having a wide geographic distribution, are much less voluminous and have variations in $^{206}Pb/^{207}Pb$, suggesting separation of sources for 1.5 b.y. or more (Sun and Hanson, 1975). These source regions are enriched in large-ion lithophile (LIL) elements and either are more variable in composition than MORB source regions or are mixed with variable amounts of MORB before they reach the surface.

The leading question for long term evolution of the mantle might be said to be its heterogeneity; in particular, the extent to which it is radial: a chemically layered mantle; or lateral: isolated blobs and regional inhomogeneities. Particularly pertinent to basalts is the history of their immediate parent materials. In the case of oceanic tholeiites, the question is whether the picritic material from which it is generated at 30–70 kilometers depth (see section 3.3.7) is itself a recent differentiate from a more ultramafic source, or a much earlier differentiate. Recent differentiation requires either remixing of previous differentiates or retention of primordial material (despite the amount of previous nearsurface differentiation probably being sufficient to utilize an entire mantle volume of rock). Differentiation long ago requires an intervening storage place for this picritic eclogite, and therefore, dense material.

Schilling (1971) and Sun and Hanson (1975) proposed that the low-velocity zone, or asthenosphere, is the depleted reservoir and the source of mid-ocean ridge basalts. Plume basalts, i.e., magmas from the nondepleted reservoir, are then attributed to deeper sources. A difficulty of this hypothesis is that the volume of MORB, integrated over time, is greater than the volume of the low velocity zone. As an alternative to deep sources of plume basalts, Tatsumoto (1978), Hedge (1978) and Frey *et al.* (1978) proposed that the enriched reservoir is shallow and the depleted MORB reservoir is deeper. The requirement of basal, rather than internal, heating for convective plumes also suggests a shallow origin for plume basalts.

Another hypothesis is that differentiation during Archean led to two layers in the upper mantle (Anderson, 1979). As the Earth cooled, the base of this original crust transformed to eclogite and sank through the upper mantle, resulting in a layer of peridotite overlying a layer of picritic eclogite. Melting in the eclogite layer then leads to ascent of the MORB material, plus a possibly pyroxenite residue underplating of older oceanic lithosphere. The basalt and eclogite portions of the oceanic lithosphere return to the eclogite layer by subduction. The shallow refractory peridotite layer, enriched by metasomatism, is hypothesized to be the locus of the source regions for the less abundant continental and ocean island basalts. In this hypothesis the picritic eclogite diapirs become buoyant upon partial melting and melt extensively upon ascent. This evolution is possible because eclogite has a small melting interval. Nearly complete melting of the eclogite source material is required to satisfy the bulk chemistry and trace element uniformity of MORB. The feasible degree of melting upon pressure release and the rate of rise of a diapir relative to its heat loss rate both need better quantification.

Composition of the mantle to 670 km depth

It is usually inferred that olivine is the main constituent of the mantle. This inference is based primarily on petrological evidence but it is also consistent with the seismic data. This evidence applies most strongly to the region of the mantle above about 220 km. Between 220 and 670 km deep, the seismic properties are consistent with either garnet-rich peridotite or eclogite.

Ringwood (1975) and others hypothesize that the refractory peridotite in the uppermost mantle grades into a mantle material that could be the parent (or grandparent) material of MORB: pyrolite, compositionally equivalent to a garnet-rich peridotite. Such material would not yield a sharp discontinuity at depth 220 km; instead, the olivine and pyroxene would undergo transitions to spinel + stishovite + periclase forms at depths of about 350-450 km. Pyroxene and garnet react between 200 and 400 km to form a complex solid solution. As discussed in section 3.3, oceanic tholeiites have a composition which agrees experimentally with a source region that is mainly olivine and pyroxene, with a minor portion of garnet material. Xenoliths found in alkali basalts and in kimberlite pipes which tap ~200 km deep in the mantle are predominantly peridotitic, with only a minor eclogitic portion. These xenoliths are enriched in LIL and H_2O and are the potential source materials for continental tholeiites. MORB is less enriched in LIL and appears to come from a relatively dry source. Continental and ocean island basalts have a range of compositions indicating varying degrees of partial melting, but MORB is invariably tholeiitic.

The Anderson (1979) proposal that the region of the mantle between 220 and 670 km is eclogitic is based mainly on the sharpness of these discontinuities and the changes in seismicity patterns associated with them. The 220-km discontinuity is close to the base of the low-velocity zone. It is near the maximum depth of earthquakes in continental regions and in regions where young, <50 m.y.-old oceanic lithosphere has subducted. The 670-km discontinuity is also sharp and global and is close to the maximum depth of deep focus earthquakes, which are restricted to regions where old oceanic lithosphere has subducted. Below 670 km the olivine-pyroxene material goes to ilmenite and perovskite structures which are denser than the garnet phases.

Composition of the mantle below 670 km

Olivine in the denser ilmenite and perovskite phases as the main constituent below 670 km is consistent with seismic velocities. The principal debate in interpreting seismological and high pressure laboratory data is whether the Fe:Mg ratio must increase with depth. Part of this problem is the temperature gradient. If the gradient is adiabatic, then the increase falls within the $\sim \pm$ 5% uncertainty of shock wave experiments for density of olivine material as a function of pressure and Fe:Mg ratio.

In any case, the lower mantle composition is of interest relevant to the basalt source problem because any plausible calculation of the crustal recycling rate extrapolated back in time requires a repository for returned material appreciably larger than the mantle above 670 km depth. *If* returned eclogite can separate from peridotite, then the mantle below 670 km would be largely residual, refractory peridotite. If this separation does not occur, then the lower mantle would have a pyrolitic composition depleted in some large-ion lithophiles.

While the above-stated bulk calculation applies to the problem of mantle interchange over the entire 4.5 b.y. history of the Earth, the evolution of isotopes—particularly ^{87}Sr and ^{143}Nd—constrains more strongly the evolution over the last 2.5 b.y. or so. Both isotopes require that the MORB be drawn from a reservoir depleted in these lithophiles. *If* this depletion can occur *only* by crustal differentiation, then over this time not more than one-third of the mantle can be depleted and contributing to MORB (section 7.4.2; Wasserburg and DePaolo, 1979; O'Nions *et al.*, 1979). The obvious hypothesis is that this depleted reservoir is the mantle above 670 km depth, about one-fourth of the total mass of the mantle. But to decide whether this region is the only plausible reservoir requires more careful considerations of the scenarios by which the present mantle came to be.

Evolution of the mantle

As discussed in section 9.5.1, the thermal evolution of the Earth has been most influenced by a decline in heat sources. We wish to explore what this history implies for mantle composition. As estimated in section 4.5.1, the primitive mantle had roughly the composition of an "undepleted" garnet peridotite or a peridotitic komatiite (see Tables 4.5.2-3). Early differentiation processes plausibly could have led to the development of a thick mafic-felsic crust, too buoyant to be subducted, leaving a peridotite mantle still containing an appreciable component of crustal material. As the outer layer cooled, some of the crustal mafics would have converted to eclogite. If there were large enough agglomerations, this eclogite would have settled through the upper mantle and have been halted near 670 km by the phase changes of the peridotite mantle. Partial melting would have concentrated lithophile elements into a protocontinental crust. Whether this process would have been more efficient in the Earth's history when the geothermal gradient was high is debatable. Certainly mantle heterogeneities in the Rb/Sr and Sm/Nd systems developed more strongly after Archean times (see section 7.4.2).

As discussed in section 3.3, the final fractionation leading to mid-ocean ridge basalts occurs at shallow depth, less than 80 km, enriched in plagioclase and probably depleted in olivine relative to a "pyrolite" peridotite. The parent magma is picritic (sections 3.3.7-8; O'Hara, 1968; Green *et al.*, 1979). Whether this source could be modified eclogite rather than modified peridotite depends on mechanisms of eclogite ascent, as well as the efficient global scale mixing mentioned above. If there were an eclogite layer, the highest temperatures in the mantle, relative to melting temperatures, would be toward the bottom of the thermal boundary layer at about 220 km deep. For this depth, estimates of the melting temperature of eclogite are about 200-300°C less than peridotite.

In summary, as the Earth's mantle differentiates and cools it is evolving from a more homogeneous to a more heterogeneous system. If basalts are classified according to age, then the variety of characteristics in isotopic, major, and trace element chemistry becomes more and more marked with time. For at least the last ~2.0 b.y. the isotopes demand separation of sources among continental and ocean island basalts. On the other hand, they suggest the sources of ocean ridge basalts could be globally homogeneous and well-mixed. The central question is the degree to which lateral heterogeneity has evolved into a layered mantle. Since Birch (1952), the main (but not unanimous) push has been to reconcile the seismological and equation-of-state data with a uniform composition, accounting for discontinuities with phase transitions. If the 220 km discontinuity is sharp, however, there must be some compositional change, since an appropriate phase transition of abundant material is lacking. As high pressure experimental and seismological data are refined, similar difficulties may evolve in reconciling the seismic velocity curve below 600 km with compositional homogeneity.

10.3.2 The Moon

The principal relevant properties of the Moon are conveniently characterized as geological, compositional and geophysical. In the following discussion, we place emphasis on differences from the Earth.

Geological properties

The dominant shaper of the surface is exogenic: the great basins and their consequent throwout and stress patterns, as discussed in sections 5.4 and 6.5. Some of the manifestations of internally generated activity, such as sinuous rilles, appear to be influenced in their occurrence by the basins. Probably the most important geological constraints in lunar evolution over the last ~4.4 b.y. are: (1) the net amount of expansion or contraction since a single coherent surface was established is very slight, less than ~0.1%; (2) morphometry and remote sensing of ancient basins and upland plains indicate voluminous volcanism from shallow sources; (3) the post-Imbrian mare lava flows constitute a very minor proportion of the total crust, perhaps 0.2% of the total volume (Hörz, 1978). These properties indicate that the main course of lunar history has been one of activity declining to negligible levels. The lack of distinct overall expansion or contraction requires compensatory changes within the volume integral of temperature over the applicable time interval (Solomon and Chaiken, 1976; see also section 9.5.2). The only way such a balance could plausibly occur would be for the outer part of the Moon to be cooling as the center heats up, implying an ever thickening lithosphere. There is some uncertainty as to the effective time of establishment of a "single coherent surface," but it is not more recent than 3.9 b.y. ago; i.e., not since the Imbrium and Orientale impacts. An auxiliary indicator of a predominantly cooling regime is younger ages for compressive features (wrinkle ridges) than tensile (circum-maria rilles) in ringed maria (Solomon and Head, 1979).

The small volume of mare volcanism, its severe partitioning of trace elements, and its occurrence 0.5–1.2 b.y. after differentiation of the bulk of the crust imply rather deep source regions. The low viscosity implied by the great extent and low scarp heights of Imbrium flows (Schaber, 1973) also suggests melting of deep source regions undersaturated in silica.

Compositional properties

Lunar rocks, as discussed in sections 1.2.9–10, 2.2.1 and 7.3 have several major element characteristics unmistakenly indicative of a constitution and history quite different from the Earth's:

A dearth of volatiles: no perceptible H_2O, and lower levels of K_2O and Na_2O.

Lower oxidation level: no Fe_2O_3, in addition to several lesser indicators, such as Cr^{2+}.

Lower silica content: SiO_2 often less than 48%.

Higher content of refractory lithophiles, such as Al_2O_3.

Higher ultramafic components: FeO, MgO, and (in most) TiO_2.

These characteristics require the source regions (and hence probably the bulk of the outer half of the Moon) to be low in volatiles and possibly high in refractory silicates compared to the Earth, and the crustal differentiation to be less thorough: less reworking or hydration to obtain highly siliceous rock.

Trace element abundances and isotopic ratios also require a dry, oxygen-poor initial constitution subjected to very early major differentiation, followed by a history of markedly dwindling activity. The early differentiation extended rather deep—perhaps 300 km—since features of the mare basalts which came to the surface 3.1–3.9 b.y. ago require that either their source regions were subject to differentiation, or that the magmas had major interaction with differentiated material in their ascents. Most restrictive on lunar history is the isotopic evidence discussed in section 7.3. The Rb-Sr model age from lunar soils indicates a formation time of 4.60–4.65 b.y. ago, with an initial strontium intercept of 0.6989–0.6990. The ^{207}Pb-^{206}Pb isochron for terra rocks appears to require a differentiation which is early—4.42 b.y. ago—but distinctly later than formation. The enrichment factors of Rb-Sr for a variety of rocks and other considerations, such as the composition of KREEP basalts, indicate significant intracrustal layering at this early stage: a highly anorthositic surface layer, a more basic lower layer, and intervening zones of residual melt greatly enriched in incompatible elements. Subsequent differentiation was at a considerably lower rate, but not completely negligible, as indicated by Rb-Sr enrichment factors.

Geophysical properties

The Moon appears to be closer to isostatic equilibrium than the Earth, in the sense of the stresses implied: the topographic and gravitational irregularities of the Moon are appreciably less than six times the Earth's. Assuming isostatic compensation by variation in crustal thickness and using altimetry and gravimetry to extrapolate from the seismologically determined crustal thickness in the northeast Procellarum region, one obtains ~70 km for the global mean crustal thickness. It is entirely plausible, however, that a significant part of the compensation is by lateral variations in the upper mantle Mg:Fe ratio (Wasson and Warren, 1980; section 4.2).

Travel times in the eastern Procellarum area indicate a gradual increase in seismic velocity with depth to ~6.8 km/sec at 20 km, relative constancy to 56 km depth and then a jump to more than 8 km/sec. The increase in the top 20 km is most plausibly attributed to a decrease in porosity. The 6.8 km/sec velocity in the lower crust is the principal seismological puzzle; it is difficult to reconcile with a specific anorthositic-gabbroic mix (Liebermann and Ringwood, 1976). The upper mantle velocity below 56 km is consistent with plausible olivine-pyroxene compositions.

A significant constraint on the nature of basaltic deposits on the Moon is the nature of the gravity profiles from orbits crossing ringed maria "mascons." These profiles have appreciable short wavelength content, necessarily the "edge" effects of a shallow high density layer about 1 km thick.

Summary

Basaltic volcanism on the Moon thus can be characterized as two main types: "primordial" and "evolutionary," the former associated with the early crustal differentiation $\gtrsim$4.4 b.y. ago and the latter with subsequent development.

No pristine samples certain to be primordial basalts have been obtained. There is much evidence, however, of an early differentiation of large enough scale to have had significant extrusions. While there are few rock specimens as well preserved as the ancient meteorites, the Moon constitutes the best constrained instance of very early circumstances in the inner solar system. The highland basalts indicate that, even on a satellite, conditions in the first 0.2 b.y. were hot enough to lead to major differentiations and redifferentiations. The thickness of the lunar crust requires that this heating and differentiation extend some hundreds of kilometers deep.

In contrast, the evolutionary basalts of the Moon—the mare basalts—are, on the one hand, the best documented series of rocks extant, but, on the other, representative of a quantitatively minor process. The mare basalts are only a fraction of a percent of the bulk of the lunar crust. However, they indicate that partial melting and the resultant volcanism are rather inexorable processes. Even on a very dry smallish planet which has already undergone a significant differentiation, magmatism at appreciable depth can produce basalts with enough pressure behind them to come to the surface, so long as the planet is not strongly cooling overall. Furthermore, such basalts can have some variety of compositions, with trends which are not entirely associated with temporal evolution, indicating heterogeneity of source regions.

The Moon is undoubtedly the best understood planet in its evolution over the last ~4.4 b.y. However, the Moon's evolution (as distinguished from origin) is still not without mysteries. In particular, the source of the fields creating the remanent magnetism as young as 3.5 b.y. is obscure (but see Stevenson, 1980).

10.3.3 Mars

Spectacular volcanoes, up to 27 km high with flows up to 800 km in length, together with extensive plains flows, provide clear evidence of the importance of volcanism on Mars. Low impact crater counts have been found for some areas, requiring activity within the last 1 b.y. on even the longest time scales, and within the last ~200 m.y. on the shortest. Estimates of the predominant ages on the Tharsis shield volcanoes range from about 500 m.y. to more than 3.0 b.y. (see Fig. 8.10.1). However, while the period of peak activity is uncertain, the duration of martian volcanism and associated tectonics is very long.

There is abundant and varied evidence suggesting that the volcanism is dominantly basaltic. As described in section 2.2.5, Earth-based reflectance spectra of the dark areas have long been interpreted as being consistent with a basaltic mineralogy. This interpretation is not based upon curve shape alone but upon the observation of at least one absorption band at ~1.0 μm, suggestive of octahedrally coordinated Fe^{2+} as in olivine and/or pyroxene. The bright area spectra also have sometimes been interpreted as containing a "surviving" pyroxene component and magnetite, although the spectra of these areas are generally thought to be dominated by weathered materials such as clays, hydrated oxides of iron and possibly salts. Bright region spectra are nearly identical over the planet, while significant differences occur among dark region spectra. A weathered "bright spectrum" assemblage is generally derivable from mafic rocks by processes thought to occur under Mars's surface (and possibly subsurface) conditions, but it is difficult to "work back" to a primary igneous rock composition by consideration of weathering products alone (see below).

The Viking lander sites, which may be more boulder-rich than the average of Mars, show an abundance of extensively vesiculated boulders which have a different reflectance spectra from the dust. On the basis of spectral reflectance measurements by the lander imagery system, the soil appears to be an assemblage of weathering products—especially Fe-rich clays. However, the boulders seem to have low albedos and a visible reflectance compatible with a possible basaltic composition. The lander X-ray fluorescence analysis experiment (XRF), a partial elemental analysis, was found to be compatible with an assemblage of weathering products dominated by Fe-rich clay, such as is suggested by the lander spectral results. One question is whether the weathering products must represent a "closed-system" alteration of the source igneous rocks, or whether the surface weathering products could have been fractionated and segregated by winds or other agents. On Earth, weathering products tend to become chemically and mechanically segregated into sandstones, limestones, shales and evaporites, but in the absence of liquid water, the same may not be true for Mars. If so, the XRF experiment results on the weathered soil could correspond to that of the original igneous rock, which would be unlike any meteorite (except possibly shergottites) or terrestrial rock, but more similar to primitive basalts or some ultramafics than to other types. For example, the SiO_2 content of the soils is $\leq$50%, which suggests a basaltic or even more mafic rock. Thus the bulk of the Earth-based and Viking lander data favor a basaltic or ultramafic mineralogy.

Also, the great length of lava flows (up to 800 km) and the low slopes (4–6°) suggest not only high effusion rates, but also low viscosity, low yield strength magmas; i.e., iron-rich basaltic or ultramafic magmas. There seem to be fewer spatter cones or side vents on Mars relative to the Earth, consistent with a fluid magma. There is some complexity involved in such interpretations, however, owing to the importance of eruption *rates* in determining flow lengths, slopes, etc., but even a high rate suggests low viscosity. Therefore, morphological observations do support a mafic/ultramafic composition for eruptive rocks on Mars, but are slightly more ambiguous than spectroscopic observations. In summary, volcanic activity has been a dominant force in supplying material to the surface of Mars and creating the surface morphology as well.

There evidently has been extensive and prolonged mafic/ultramafic volcanism on Mars. Two factors are key to rationalizing such activity: (1) bulk planetary (or bulk mantle) chemical composition, and (2) planetary energy history, as reflected in time-temperature curves for each depth, especially those depths which might participate in supplying surface magmas. Mars either was hot at one time or accreted in a partially inhomogeneous way because it definitely has a substantial (1500–2000 km) core much denser than Mars's average density. Knowledge of the radial density distribution is too rudimentary to allow us to say what the proportion of Fe^0 to FeS to FeO might be in Mars's core, but at least one is likely to be an important constituent. A core would satisfy the density and moment-of-inertia constraints while yielding a thermal evolution consistent with the tectonic history. Most of the tectonic features on Mars appear to be extensional (see section 6.5), which argues against large scale cooling over the last 1–2 b.y. of Mars's history. The relative youth of Olympus Mons—some almost uncratered areas on its flanks—also suggests that Mars stayed warm rather late.

The absence of marked cooling in the last 2.0 b.y. precludes drastic upward differentiation of lithophilic radioactive elements. If there had been such differentiation, then the planet would have been cooling considerably during the last 2 b.y. and the extensional features, coupled with the apparent absence of compressional features, would not be evident.

The evidence is sketchy as to whether Mars was significantly heated early in its evolution. Erosional effects, such as dust deposition, have modified or obliterated many features (see section 8.7). There is an indication that formation of a chemically distinct martian crust occurred prior to 3.5 b.y. ago, in that old terrain features, such as Hellas, appear compensated. Therefore, fairly shallow vertical density differences were set in below at some time. If they were set in *after* Hellas formed, then Hellas's shape would have been destroyed at the time of differentiation. On the other hand, it is possible that compensation could have been achieved without much surface distortion. Also much smaller bodies—Mercury, the Moon and the eucrite parent body—differentiated crusts very early, and hence to assume that Mars experienced a similar early burst of energy seems less *ad hoc* than to assume it somehow escaped (see section 9.5.3).

As discussed in sections 4.3 and 10.2, the dryness of Mars poses problems for nebula models and the resulting planetary compositions. In any case, the apparent dearth of volatiles, the modest degree of central concentration required by the moment-of-inertia, the mean density, and theoretically deduced temperature-depth curves produce a Mars mantle model with a high FeO content (~21%). Despite uncertainties in $FeO/FeS/Fe^0$ ratios and core composition, Mars's mantle must be denser, uncompressed, than the Earth's. Almost certainly this is due to a higher FeO content, but the uncertainties alluded to above allow FeO to range from 14% to 30%.

The only major point of convergence of all the models is that Mars's mantle is more FeO rich by a factor of, say, two or three than that of Earth. The question is whether martian volcanism as discussed above could be reconstructed with a martian mantle that was exactly like that of the Earth in every way *except* for a much higher FeO content. Plausible Earth mantle compositions generate normative mantle mineralogies which produce cotectic-like melts like olivine tholeiites. As discussed in section 3.6, changing the composition only by adding FeO produces a different normative mantle composition for Mars. Melting of that composition produces iron-rich and ultrabasic lavas. These melts would have a composition similar to that observed by the Viking lander X-ray-fluorescence experiment. This brings us back to a question we raised earlier: "What does an assemblage of weathering products tell you about the primary igneous source from which they were derived?" On Mars, it probably tells us a great deal. It seems as if at least recent alteration of Mars took place without oceans. Ultraviolet-activated H_2O vapor-rock interaction or perhaps local subsurface liquids appear to have been responsible. In either case, the ion mobility on Mars was probably low. Moreover, the general agreement between analyses of Viking 1 and 2 lander material supports the notion that alteration on Mars was of the closed-system variety, in contrast to most sedimentary processes on Earth. Similarly, fractionations associated with the processes that produced the duricrust, or caused by aeolean transport, seem slight.

Hence there is good coupling between the observed characteristics of igneous activity on Mars, global geophysical constraints, and the high FeO content inferred from formation models. This consistent story concerning Mars volcanism can be put together without reference to the bulk volatile content of Mars, its degassing history and its atmospheric history, because, as emphasized in Chapter 3, volatiles have little influence on abundant basalts. Mars's probable accretion history, its subsequent thermal history, zonal structure, and obvious widespread volcanism all suggest that there has been ample opportunity for at least the outer portion of Mars to outgas. Hence the small Mars surface and atmospheric inventories of ^{36}Ar and other volatiles suggest that Mars is a volatile poor planet. If there was an early atmospheric loss, it is uncertain whether it would produce (an unobserved) elemental fractionation in the remaining nonradiogenic rare gas inventory. Another problem is potassium. If the Soviet Mars 5 orbital γ-ray data and interpretations are correct, then the *global* surface K/U ratio for Mars is only one-third that of the Earth's crust and one-tenth that of chondrites (Anders and Owen, 1977). Hence for any models but those for a volatile-poor Mars, it seems necessary to lose the early atmosphere, to bury much of the K in the interior, and to preserve any remaining volatiles from anything like efficient outgassing despite continuing volcanism. A volatile-poor Mars is difficult to reconcile with systematic patterns of planetary properties as the result of origin. Aside from the nonmonotonic trend in volatiles with solar distance, Mars seems to break the coupling between volatile content and degree of oxidation—a key feature in some planetary formation models.

Many of the controversies could probably be resolved by a combination of global geochemical (γ-ray) and mineralogical mapping from orbit and analysis of carefully selected samples from several sites. In the meantime, an Fe-rich martian mantle and the apparent occurrence of widespread basaltic/ultramafic volcanism on Mars are two of the very few things about the planet on which almost all investigators agree.

10.3.4 Mercury

As discussed in sections 2.2.3 and 5.5, direct evidence of basaltic composition on Mercury's surface is limited to reflection spectrophotometry indicating pyroxene, and a morphology of some units, such as the broad plains around the Caloris Basin, suggestive of low viscosity lava flows.

Indirect indicators of volcanism are those features of Mercury requiring a fairly warm thermal history. The predominantly compressive tectonics of the surface, marked by lobate scarps, suggests an early differentiation, bringing most of the radioactive lithophiles near the surface, so as to produce a net cooling over most of Mercury's evolution. The high density and the magnetic field of Mercury require a core of metallic iron, in contrast to the silicate surface. The initial high temperature of Mercury's material plus the energy released by core formation would have been sufficient for an early crustal differentiation.

There are indications of early intensive volcanism in the extensive upland "incrater" plains of Mercury (sections 8.5–7). However, for most of its history Mercury has been the most compressive in its tectonics of all the terrestrial planets (now that significant rifts have been found on Venus: Pettengill *et al.*, 1979a). As discussed in section 9.5, the onset of the net cooling to produce a compressive regime cuts off volcanism in a one-plate planet. Hence Mercury's volcanism probably terminated earlier than that of the other terrestrial planets or the Moon, perhaps not long after the "cataclysm" of 3.9 b.y. ago, of which the Caloris Basin was plausibly a product. This conjecture is not inconsistent with the cratering data: Mercury's surface is cratered everywhere, but the absolute time scale is rather uncertain (see section 8.10).

The greatest problems of explaining Mercury's constitution go back to origin circumstances (see section 4.5.3): in particular, the depletion of magnesian silicates relative to iron necessary to explain its high mean density. It is likely that the oxidation level on Mercury is sufficiently low that despite the low bulk content of magnesium, the comparatively small mantle probably has a higher Mg:Fe ratio than that of the Earth or the Moon.

The Moon is probably the best guide for the nature of basaltic rocks which may eventually be sampled on the mercurian surface. We should expect to find a highly brecciated surface with specimens that, while not identifiable as basalts petrographically, will have compositions corresponding to the basic component of a primordial crustal differentiation. The mercurian analogue of mare basalts should be scarce or entirely absent. The failure to observe dark plains is consistent with the modelling hypothesis that volcanism gets cut off early in a one-plate planet of predominantly cooling thermal history. Whether the broad plains around Caloris are some sort of lava flow—perhaps impact-triggered—rather than relatively fine ejecta, is rather conjectural. All these inferences are, of course, derived from spacecraft observations of only half the planet.

10.3.5 Venus

Radar measurements of the near equatorial topography of Venus showed variations in height of $\pm$ 3 km and wavelengths of 3000 km (Campbell *et al.*, 1972). Such large variations imply isostasy, and hence a crustal differentiation. The first positive indication of differentiation was the high content of K, U, Th inferred from the Venera 8 gamma-ray experiment (Vinogradov *et al.*, 1973). Now the Pioneer Venus radar mapper experiment shows that while most of Venus's surface is gently undulating, it has two sizeable plateaus, both of ~7 km relief (Pettengill *et al.*, 1980). Hence Venus must have been tectonically active to the extent that granite-like radioactive content, at some locations, is not implausible. The most unmistakable difference between the tectonics of Venus and the Earth is the absence of any feature like an ocean rise. Hence, if plate tectonics ever existed on Venus, it must have terminated, and the surface must be entirely choked with crust (Phillips *et al.*, 1981).

As described in section 4.5, the venusian mantle would have MgO/(MgO + FeO) ratio like the Earth's, if its lower mean density were the consequence of an iron or sulfur deficiency, but it would have a somewhat lower ratio if the lower density came from an oxygen excess. Since volatiles apparently have a negligible role in volumetrically major basaltic differentiations (see section 3.6), the abundant basalts on Venus should be similar to the abundant basalts on Earth. However, this conjecture is based upon the assumption that Venus's density has a simple explanation. At our present state of ignorance, it cannot be ruled out that Venus is both appreciably depleted in iron and has a low oxidation level: i.e., that it is somewhere between the Earth and the Moon in mantle constitution. Hence, it could conceivably have basalts intermediate between lunar and terrestrial in major element chemistry and in characteristics dependent on redox state. An intriguing question will be the degree of siderophile depletion in venusian surface rocks, since such depletion is an indicator of whether material has been in a planet which has undergone core differentiation (Ringwood, 1979).

While volatiles may not influence abundant basalt character—terrestrial tholeiites and their venusian analogues, whatever they may be—they could be significant in the more complex and later stages of crustal differentiation, such as those associated with subduction zones, continental rifting, "hot spots," etc., on Earth. It is likely that Venus's interior has a lower volatile content than the Earth. An outgassing efficiency comparable to that of the Earth probably existed in the first 2 or 3 b.y. Once the surface became hot enough for any water to boil off, it would no longer be recycled, as it is on Earth. The high surface temperature keeps the venusian lithosphere from becoming sufficiently more dense than the underlying mantle to subduct.

While Venus appears to be tectonically dead now, it could have had a rich history, of which the two plateaus are remnants. Hence a balanced exploration program requires mapping by radar satellites as well as landers to obtain compositional information directly. It would be a pity to have several landers all fall on venusian analogues of ocean floor and Precambrian shield, and miss the critical one percent of surface area which contains analogues of island arcs, rift volcanism and other important features of limited extent.

10.3.6 Galilean satellites

Asteroids and achondrite parent bodies can plausibly be passed over as nonevolving bodies (see section 10.4.2 for possible exceptions), and the outer Galilean satellites as icy rather than terrestrial planets. The same cannot be done with the dense inner Galilean satellites, Io and Europa, which now have been photographed by the Voyager probes.

Io manifestly must have an energy source different from the other terrestrial bodies. This source, probably tidal (Peale *et al.*, 1979), appears to be rather shallow, because of the widespread occurrence of volcanism: more than 300 possible volcanic vents have been found. The explosiveness of the volcanism probably arises from its volatile content; in fact, the volcanism is probably not basaltic at all, but constituted of more volatile sulfur compounds, which would drastically lower the energy requirements. Although only sulfur compounds have been identified spectroscopically, the morphology of some calderas suggests that silicate volcanism exists.

Europa also appears to be nonbasaltic in its near surface evolution. The mildness of the topographic variations observed by Voyager 2 suggests a predominantly icy surface layer, consistent with theoretical models.

10.4 LEADING INFERENCES

In this section, we will attempt to set up a general model of terrestrial planet evolution, then go back to reexamine the planets with a view to identifying variations and exceptions. Next, we will discuss the consequences of the general model and variations for basaltic volcanism. Finally, we will suggest the outstanding problems and directions for future research. Aside from earlier chapters of this book, principal references are Kaula (1975) and Walker *et al.* (1979).

10.4.1 A general model

The most obvious independent variable is planetary mass. As discussed in section 10.2.6, the more massive a planet, the more heating it is likely to have in the formation process. The absolute magnitude of this heating will depend on the efficiency of heat retention upon impact. The relative magnitude for planets formed from planetesimals appreciably smaller than themselves (but still big enough to bury significant heat upon impact) will be at radius r approximately proportional to r^2, up to a radius at which solidus temperatures are attained. Above this radius, convection should keep the temperature curve close to the solidus.

A priori, there is no clear rule as to the minimum-sized planet to approach melting temperature in its formation and thus to attain a convective regime in its outer parts. This uncertainty is primarily due to ignorance as to the efficiency of heat retention in large impacts. If we adopt h = 50% for the retention factor (Kaula, 1980a), then the minimum-sized planet to attain melting temperatures in its outer parts, determined from (cf. section 10.2.3)

$$\max(\Delta T) \approx h\,\frac{GM}{RC_p}\left(1 + \frac{1}{2\theta}\right) \approx 1700^\circ\,\mathrm{K} \qquad (10.4.1)$$

will be R ~ 1900 km, using 3.7 g-cm^{-3} for density and 4 for θ, the mean ratio GM/Rv^2, where v is approach velocity. The most important consequence of attaining such temperatures is obviously core formation, which adds an extra pulse of energy yielding an average temperature change

$$\Delta T \approx (1-P)\,P\,\Delta\rho\,\frac{GM}{3R\bar{\rho}C_p} \qquad (10.4.2)$$

where P is the fraction of the mass in the core, $\Delta\rho$ is the density difference between core and mantle, $\bar{\rho}$ is the mean density, and the other symbols have their usual meanings. For a large planet, core formation probably started well before the completion of planetary growth, but the thermal result will be roughly the same. The temperature rise in the later stages may be great enough that, with the help of instability of a lighter center (Stevenson, 1980), core formation is a rapidly self-accelerating process. An important feature of core formation is that the energy released tends to be maximized near the core-mantle boundary, thus further inducing mantle-wide convection. Core formation also has compositional implications: it would remove siderophiles and chalcophiles from the mantle, plus, in the Earth and Venus, a portion of elements which may undergo a change in character at higher pressures, such as has been suggested for potassium. For a small planet of relatively high oxidation level, core formation may be delayed until temperatures are raised radioactively, and may be a slow process because the temperature rises are modest.

Other effects of early heating will be outgassing and crustal differentiation. Outgassing will evidently go more rapidly the greater the heating. Crustal separation is a different matter, since it involves rocks of varying ratios of calc-aluminous to ferromagnesian silicates, and hence of small density differentials. The vigorous convection induced by heating and outgassing may bring about remixing rather than separation, particularly if more than a small amount of melting is involved. However, the strong inverse correlation of crustal thickness with planetary size (see Table 4.2.1) may be related to the depth at which the basalt-eclogite phase transition intersects the solidus, and hence below which calc-aluminous material is more likely to sink than rise: about 50 km on the Earth, but 300 km on the Moon. Positively correlated with planet size must be the degree of recycling of crustal material to the mantle. This process affects thermal evolution because the degree of upward differentiation of U, Th, K will be proportionate to the degree of crustal differentiation. This proportionality should be nonlinear, because large-ion lithophiles are upwardly enriched more severely and are less likely to be recycled than the major crustal constituents. In any case, it seems that an additional factor contributing to the prolongation of activity in larger planets is the return, or retention, of more radioactivity in their mantles.

Radioactive heat should become the dominant energy source within a time that the heat would melt its host rocks if it were not removed, since formation heat in excess of this amount would have largely been removed by convection. For chondritic composition this time is 1.2 b.y. The time is somewhat longer for planets with deficiencies in potassium. Therefore, it is useful to discuss idealizations of planetary evolution whose heat sources are entirely radiogenic as, indeed, was customary until rather recently.

While the ultimate asymptote of planetary evolution (ignoring solar evolution) is complete upward differentiation of heat sources which meanwhile have dwindled to negligible rates, a more useful quasi-asymptote is a planet whose heat loss rate is equal to its radioactive heat generation rate. Given a distribution of heat generation Q(r) with radius r, from the equation of heat transfer (Eq. 9.3.1), the temperature at depth $D \ll R$, the planetary radius, will be $T(D) \approx$

$$T(0) + \frac{D}{4\pi R^2 K}\int_0^R Q(r)\, r^2 dr - \frac{1}{K}\int_0^D\int_0^y Q(z)dz\, dy \quad (10.4.3)$$

where K is the thermal conductivity. With uniform heat sources, Eq. (10.4.3) predicts that the temperature rise T(D)-T(0) will vary proportionate to R. Given further the observation that small planets appear to have been more effective at upwardly differentiating lithophiles and their associated heat sources, then the proportionality will have a higher than linear dependence on R. Hence the smaller a planet, the thicker the lithosphere—the layer in which radial heat transfer is dependent on conduction rather than on convection. This condition is true for any planet which has cooled off enough that the outer layers are representable by a single continuous lithosphere, such as the Moon, Mercury, and Mars. It is partly true for planets characterized by plate tectonics, in which some heat is brought essentially all the way to the surface by convection.

The foregoing discussion of lithospheric thickness brought up the temperature dependence of viscosity in an implicit manner. It is observed that silicate viscosities tend to be proportionated to $\exp(T_m/T)$, where T_m is the melting temperature. Since the melting gradient is steeper than the adiabatic gradient, as discussed in section 9.4.4 and 9.5.1, a convecting planet may tend toward a state where the viscosity in the convecting zone will increase with depth. To some extent this increase would be counteracted by the "governor" tendencies of temperature-dependent viscosity: the higher the temperature in a region, the quicker convection can cool off the region by removing heat, and vice-versa. But it seems likely a planet will develop a zone of high viscosity toward the bottom of the mantle. An important unknown affecting this inference is the pressure dependence of viscosity (Kaula, 1980b).

The transition of a planet from the formation-dominated early hot stages to the radioactivity-dominated late quasi-equilibrium stages will depend on a variety of factors. A body which is small, and therefore of high area:mass ratio, can get rid of its heat quickly and hence approach equilibrium more rapidly. For bodies of less than 500 km radius, such as the asteroids and the meteorite parent bodies, the characteristic time of this temperature decay is short compared to the ~1.2 b.y. of the radioactive heating time. Therefore, these bodies are dismissed in section 10.3 as "nonevolving." For a body of ~1700 km radius, such as the Moon, the situation is marginal; consequently, there is an era of interaction between cooling of the outer parts and heating of the inner parts by radioactivity before the cooling prevails to the extent that an impenetrably thick lithospheric layer dominates the transfer of both heat and magma. For a body of large radius, say more than 5000 km, the much greater store of initial energy, the low area:mass ratio, and the relative inefficiency of upward differentiation of large-ion lithophiles keep it active until current times.

The preceding discussion is based almost entirely on one independent variable: the mass. Next in significance is bulk composition. Composition in turn depends primarily on distance from the sun and secondarily on planetary mass: there appear to be positive correlations of volatile content and oxidation level with distance from the sun and planet size. The correlation with solar distance is obviously a consequence of a negative temperature gradient in the nebula until gases were expelled, as well as planets being made from material in their respective zones (see section 10.2). The correlation with planetary mass does not have as clear an explanation, the most plausible being an addition of more volatile material to the inner nebula late during planetary growth which the larger planets acquired more easily. This hypothesis does not, however, explain the remarkable differences in inert gas abundances among Venus, the Earth and Mars.

The most evident consequence of compositional difference is the contrast in evolution of Mercury and Mars. The planet of low oxidation and volatile level, Mercury, had a large component of reduced iron and differentiated a core early, thus attaining a net cooling and a compressive tectonic regime for most of its history. The planet of high oxidation level, Mars, had a marginally small amount of reduced iron, and hence apparently required radiogenic heat to attain core separation circumstances relatively late in its evolution. However, these thermal circumstances led to longer persistence of a tensile tectonic regime, so that the major volcanic constructs of Tharsis could occur.

A third variable affecting planetary evolution is occurrence as a satellite rather than as a planet. In general, a satellite should reflect hotter circumstances of formation than an isolated body of the same size, due to the influence of its planetary companion. The Moon and Galilean satellites indicate that composition is greatly affected. Planetary influence may also be felt in the course of evolution, as is most evident for Io.

A fourth variable is surface temperature. While it may have little influence on bulk evolution because of the temperature dependence of viscosity, it may be of significance for near surface differentiations through its influence on magma thermodynamics, buoyancy of the lithosphere, and tectonic settings.

Finally, volatile content as a variable independent from mass is important for crustal evolution for its influence on the character of volcanism and subduction (and hence recycling), and on more advanced differentiations, andesitic to granitic.

In summary, the larger a planet, the hotter it runs; however, the increase in heat transport effectiveness with temperature is such that a terrestrial planet should rarely get much above melting for realistic growth times. The larger a planet, the longer it takes to attain a quasi-equilibrium state, in which the heat loss is not much more than contemporaneous radioactive heat generation. Also positively correlated with planet size are complexity and variety, arising from, first, the domination of convection in heat transfer, leading to a plate tectonic surface situation; second, the greater degree of recycling of crustal material; third, the attainment of pressures sufficient for more phase transitions, which have strong bearing on both heat transport and differentiation; and fourth, the higher volatile content.

10.4.2 Variations and exceptions

We list here phenomena unexplained by, or difficult to reconcile with, the general picture we have presented:

(1) The low Mg:Fe ratio in Mercury. As discussed in section 4.5.3, there does exist a narrow temperature band in which iron condenses but magnesian silicates do not; however, this seems too weak an effect to account for the large departure of Mercury from the chondritic ratio in a nebula of realistically evolving temperature. Consequently, resort must be made to effects dependent on dynamical evolution, as discussed in section 4.5.

(2) The lower density of Venus compared to the Earth. Explanation of the difference by a lower sulfur content in Venus but a solar complement in the Earth is inconsistent with depletion of other volatiles in the Earth. The alternate explanations by higher oxidation or lower iron content in Venus are inconsistent with monotonic trends in nebula properties, and require resort to dynamical effects, perhaps associated with Venus's lack of a satellite, as discussed in section 4.5. The explanation by a thick basaltic, rather than eclogitic, layer requires strong upward differentiation of lithophiles and higher temperatures at depth (to keep

basalt above the solidus at pressures greater than ~15kb) to be quantitatively sufficient.

(3) The higher inert gas content of Venus. Hypothesis dependent on steep radial gradients at, or soon after, condensation in the nebula (e.g., solar wind implantation or pressure effects) severely limit the allowable subsequent interzonal mixing to account for the order-of-magnitude differences among Venus, Earth and Mars in abundance of primordial argon. Hence acquisition of primordial argon by large body impact in the late stages of formation seems more plausible.

(4) The drastic differences in evolution of Venus and the Earth: a water content smaller by a factor of more than 10^{-2}, a magnetic field smaller by a factor of more than 10^{-4}. Neat explanations exist for both of these. For the water content, Venus is close enough to the sun for the water to be boiled off and dissociated (Walker, 1975; Pollack, 1979). For the magnetic field, Venus lacks an energy source from either precessional torques—no Moon—or core differentiation, since its central pressure is too low (Stevenson *et al.*, 1980). However, both explanations are only marginally satisfying, leading to speculation as to whether Venus was different in the past.

(5) Evidence of mantle heterogeneity. Departures from spherical symmetry and compositional homogeneity must exist to some degree in all the terrestrial planet mantles. However, the Earth dominates our attention not only because of accessibility, but also because of the greater range of conditions. The most marked heterogeneity in the Earth is the simultaneous presence of a reduced iron core and volatiles in the outer parts. Also rather inescapable is the need to keep the source regions of certain categories of rocks, such as basalts from continents and ocean rises separated for more than ~2.0 b.y. as discussed in section 7.4. Geophysical data, such as heat flow, gravity variations, seismic velocities and the tectonic plate motions, indicate that there are significant lateral variations to a few hundred kilometers deep which must be at least partly thermal, but may also be compositional. More in debate is the need for compositional differences with depth to satisfy seismological models associated with both the 220-km and 670-km discontinuities and the deeper mantle. As discussed in section 4.5, there are differing interpretations of equations-of-state and phase transitions as to whether there is a distinct increase in Fe/Mg and/or Si/Mg with depth below 670 km (Ringwood, 1975; Anderson, 1977). As discussed in section 10.3.1, if the 220-km discontinuity is sharp, there are no phase changes in the abundant Earth materials to match it. While it is the current general opinion that there is heat transfer throughout the mantle by convection, it is debatable whether there is transfer of matter as well as heat between lower and upper mantles (section 9.5.1; Stevenson and Turner, 1979). The principal need for all-mantle convection cells is mainly to get flow patterns consistent with plate velocities and subduction zone dip angles (Hager and O'Connell, 1979). Finally, the implications of plausible cosmogonies for mantle heterogeneity are not at all clear. While classical heterogeneous accretion, in which the entire core is formed before the mantle is added, is virtually impossible to reconcile with physically realistic models, the mantle still could be significantly influenced by smaller scale heterogeneities associated with the planetesimals that formed the Earth.

(6) The Moon's bulk composition. As discussed in section 4.5.2, the Moon definitely has drastic depletions in iron and volatiles; it may also have an enrichment in calc-aluminous silicates relative to ferromagnesian silicates. This pattern is quite inconsistent with a monotonic trend with respect to condensation temperature, thus suggesting a combination of differentiation in an earlier body or bodies together with dynamical effects, such as hitting the Earth with an extra large planetesimal (Kaula, 1977; Ringwood, 1979).

(7) The low volatile content of Mars compared to the Earth and Venus. It is difficult to account for the present atmospheric consititution of Mars, particularly the isotope ratios, without inferring a bulk composition deficient in volatiles (Anders and Owen, 1977). If Mars has a cosmic Fe:Mg ratio, the lower mean density requires a higher oxygen abundance, resulting in a smaller core and a lower Mg/(Mg + Fe) ratio in the mantle. In combining a small core and dryness, Mars resembles the Moon, and both bodies reinforce the hypothesis that the planets acquired their chemically active volatiles from an additional component similar in composition to carbonaceous chondrites rather late in the formation process, so that planet size was a significant factor in determining the amount acquired (Anders, 1977; Ringwood, 1979). The findings as to venusian composition emphasize that the explanation for the abundances of inert gases is not necessarily the same as for the abundances of active gases.

(8) The differences in composition within the asteroid belt. Most asteroids have spectra similar to carbonaceous chondrites (section 2.2.6; Chapman, 1979), which reinforces the idea that volatile content increases with distance from the sun. The minority of asteroids with spectra suggesting differentiation are moderately concentrated toward the inner edge of the

belt, and have a weak correlation with size. These statistics suggest that effects which would be somewhat random in their influence (most obviously, collisions) have some influence on the degree of differentiation of asteroids, as well as original heliocentric distance.

(9) The energy source for melting in the parent bodies of differentiated meteorites. This problem overlaps, and could be combined with, item (8) above. As discussed in section 10.2, currently it is quite speculative as to whether this source is collisional, electromagnetic, or short-lived radioactivity.

(10) The presence of relatively young achondrites. As discussed in Chapter 3 and Walker *et al.* (1979), there exist igneous meteorites of rather young age: the shergottites, about 0.6 b.y. old, and the nakhlites and chassignites, about 1.3 b.y. old. Both are rather volatile-rich cumulates. Their ages and petrology suggest they are products of differentiation on a sizeable planet. Given the limited availability of transfer mechanisms over the last ~4 b.y., the obvious scenario to produce these meteorites involves an asteroid impact on Mars, to which the obvious objection is that the resulting textures should be highly shocked, and that there should occur abundant high pressure phases. Shergottites contain maskelynite: shock-vitrified plagioclase. The physics of large impacts is imperfectly known; the scaling up from small events is not straightforward. Furthermore, the presence of volatiles (such as permafrost) greatly enhances ejection capability with little shock (Kieffer and Simonds, 1980). Hence it cannot be ruled out that moderately shocked meteorites could be ejecta from Mars (Wasson and Wetherill, 1979). Otherwise, we are left with a similar problem to item (9) above, only worse (see also section 9.5.6).

Other variations and exceptions could be selected, of course. The preceding list is to some extent a mixture of real variations from the model based on planetary size and distance from the sun, items (2), (3), (4), (6) and (7); related matters which involve more detailed consideration, items (1) and (5); or which affect only a minor portion of terrestrial matter, items (8), (9) and (10). Most of them involve solar system origin, which must have entailed a complex intermediate stage of relatively few sizeable bodies, leading to an appreciable variance of the actual outcome from the average of all possible planetary systems. For the purposes of interpreting basaltic volcanism, the terrestrial bodies must therefore be taken as having some systematic patterns, but also in some ways as having been formed differently.

10.4.3 Consequences for, and constraints from, basaltic volcanism

Ideally we should take models of planetary composition (Chapter 4) and evolution (Chapter 9) to predict basaltic volcanism properties for comparison with observation. Most important is to infer source characteristics (Chapter 3) from compositions (Chapters 1 and 2), abundance and morphology (Chapter 5), locus and manner of occurrence (Chapter 6), and chronology (Chapters 7 and 8). The discrepancies between observation and prediction should then be used to decide the next steps of data acquisition, experimentation, and modelling.

Practically, any discussion becomes shaped by those bodies from which samples have been obtained: the Earth, the Moon, and the basaltic achondrite parent body (or bodies), since the strongest constraints come from the petrological experiments discussed in Chapter 3 and the isotopic measurements discussed in Chapter 7. Hence a more sensible course is to take a combination of the strong observational constraints with the modelling considerations to predict circumstances which are less well known: surface compositions of unsampled planets; the constitution and thermal states of planetary interiors; and the past evolutions of the planets, with emphasis on their surface manifestations.

From the terrestrial, lunar, and achondritic samples we have evidence of two major phases of basaltic volcanism, which may be described as **primordial** and **evolutionary**.

Primordial basalts

Achondritic meteorites and lunar highland rocks indicate that differentiations producing basaltic compositions occurred more than 4.4 b.y. ago. This is early enough that we should expect the energy sources to be other than long-lived radioactive, and we should hope the source material to be truly primary. The eucritic meteorites show most clearly a primary source which could be chondritic in major element composition (section 3.5 and Stolper, 1977). They also suggest that magmatic differentiation in the presence of metal efficiently removes siderophiles, that the eucrite parent body (EPB) was very dry, and that heat sources much greater than accretionary existed in some small bodies.

The lunar evidence is more indirect. The Fra Mauro basalts are highly-processed fragments which apparently were a minor but significant part of the general differentiation of a crust of predominantly feldspathic composition. This differentiation was sufficiently complex that a primordial source composition is not inferrable. More important, the petrology and isotopic systematics of highland rocks, together with geophysical data, indicate that the Moon may have differentiated a primordial crust thick enough that either at least half of the Moon's bulk was involved or that the Moon is enriched in CaO and Al_2O_3 relative to chondrites. In either case, the outer part of the early Moon was extremely hot as well as extremely dry. Lateral variations in highland petrology—e.g., the distribution of KREEP—are consistent with the hypothesis that the heating came from infall of sizeable bodies.

Since the Moon is special because it is a satellite and the EPB may differ because it accreted early enough to be heated by Al^{26} or electromagnetic effects, it is difficult to extrapolate to primordial crustal formation on the terrestrial planets proper. As indicated in Table 4.2.1, there appears to be an inverse correlation of crustal thickness with size. As mentioned in section 10.2.7, this may have some relation to the basalt-eclogite phase transition. A further inference is that the larger a planet, the more it recycled its crustal material. This is obviously true for the Earth, on which no rock ages older than 3.8 b.y. have been found, but rather unsure for Venus. Whether it is true for the smaller planets, Mars and Mercury, is uncertain to the extent that the Moon may be anomalous in its early history because it is a satellite. The implication for subsequent planetary evolution is that "primordial" (i.e., resulting from formation processes) mantle material is likely to be exceptional, since any sort of crustal differentiation and recycling is bound to have different effects on some constituents than on others. This is even more true for planets which underwent early core differentiation.

While it is highly desirable to find primordial basalts on Mars or Mercury, we should not expect to find them on any body larger than Ceres (which may not have any basalt at all).

Evolutionary basalts

Terrestrial basalts and lunar mare basalts further indicate that basaltic volcanism can originate from appreciably modified sources over a long period of planetary evolution. The contrasts between lunar mare basalts and the most abundant terrestrial basalts, ocean rise tholeiites, emphasize the effects of planet size. Mare basalts are generally considered to be partial melts from depths greater than 200 km; they are either from cumulate sources or have interacted significantly with cumulates, and have severe trace element fractionations suggesting small percent partial melts or other complexities. On the other hand, ocean rise tholeiites are large percentage (~20%) partial melts which have differentiated at shallow depths, 30–50 km, either from mantle material depleted in the aluminous phases or from a more basic magma (e.g., a picrite), which has risen from a deeper differentiation. The depths of sources, inferred from high pressure phase equilibria studies, are suggestive of lithospheric thickness in the cases of both lunar mare basalts 3.1–3.9 b.y. ago and recent oceanic tholeiites. Both also show little influence of volatiles, although for different reasons: mare basalts, because of a bulk deficiency; tholeiites, because they result from large percentages of partial melting.

Both lunar and terrestrial evolutionary basalts have variations in composition with location which must at least in part be attributed to variations in source composition. The lunar variations are comparatively modest. At one time, the principal trend appeared to be an inverse correlation of Ti content with age. More recently, exceptions to this trend have been discovered. The terrestrial variations described in Chapters 1 and 3 are considerably greater, and of more interest as furnishing examples of varying conditions of basalt formation: pressure, percent melt, H_2O, CO_2, Fe/Mg, etc. Furthermore, some of these petrogenetic variations have correlations with age or with trace element constraints on source isolation. As discussed in section 7.4, some continental basalts have nearly chondritic $^{143}Nd/^{144}Nd$, which requires separation of their source from the depleted ocean rise basalt source for $\gtrsim$2.0 b.y. Although the nontholeiitic varieties are of minor abundance on Earth, some of them may be of closer resemblance to basalts on planets which are one-plate or of higher volatile content.

Combining the trends of the general model set up in section 10.4.1 with the associated basaltic characteristics outlined above, and taking into account the variations and exceptions described in section 10.4.2, we can conjecture the pattern in time of volcanic activity on the terrestrial bodies.

Mercury, being formed hot, probably differentiated both its core and its crust fairly early. Analogy with the Moon suggests that although it took longer to cool in the outer parts, the differentiation was fairly thorough. Whether it resulted in a mainly olivine mantle and a thick crust or an olivine + spinel + diopside (or melilite) mantle with a thin crust depends on a prime unknown: the degree to which Mercury's bulk composition is depleted in SiO_2 relative to MgO (Smith, 1979; Stolper, 1980). Subsequent volcanism was slight, and probably terminated before 3.5 b.y. ago because of the early onset of a compressive regime. Sampling of the mercurian surface should obtain few pristine specimens; mostly brecciated rocks of olivine tholeiite, or perhaps even komatiite-like composition with low Fe/Mg and no alkalis. Volumetrically minor lavas of less than 4.0 b.y. age may be found, with characteristics indicating formation as small percentage partial melts at considerable pressure.

Venus appears to have been tectonically active, from its altimetry (Pettengill *et al.*, 1980) and gravity field (Sjogren *et al.*, 1980). Since Venus no doubt has had as long a history as the Earth, evidence of a primordial differentiation will be obliterated and a comparable complexity should prevail. There should be about the same abundance of volcanism, and the abundant basalts should be similar to those on Earth, olivine tholeiites consistent with near surface crystal-liquid equilibrium (at, if anything, lower pressures than of Earth, because of higher surface temperature), with perhaps a higher Fe:Mg ratio on the average. Trace element contents in abundant basalts may reflect the lower pressure or differences in compositional evolution. More marked differences between Venus and the Earth in major element chemistry should occur in its less abundant basalts, such as the venusian analogues of island arc basalts, if they exist: differences in near surface temperatures and volatile content could be much more critical here. At this point it is rather speculative as to whether venusian basalts will reflect lower volatile or alkali content. A respect in which Venus may differ significantly from the Earth is the extent to which the record of the past has been obliterated by erosion.

There is still ample room for speculation about the Earth and Moon, in the attempts to extrapolate into the interior or into the past. However, the extrapolation has a different character pertaining much more to scientific inference than to what an exploring spacecraft may find. Hence we have put these questions in the next section on future directions.

Mars supports the highest pile of basalt in the solar system, possibly because it still maintained a tensional tectonic state after it had already developed a rather thick lithosphere. This later-stage tensional state was plausibly induced by later core formation, in turn the consequence of the relatively small portion of free iron. The height and spacing of the volcanoes are dependent on both the thickness and the depth of sources. Source depths are in turn proportionate to lithospheric thickness, and hence to mechanical strength. Therefore, the pressures reflected by the younger martian basalt compositions may be even higher than those indicated by mare basalts, a consequence of the higher gravity and the longer persistence of tensional tectonics. The principal effect of the higher oxidation level should be a higher Fe:Mg ratio (McGetchin and Smyth, 1978). The low viscosity suggested by the low slopes of lava flows is consistent with this higher iron content. More uncertain is whether there is a higher abundance than in the Earth of elements of intermediate volatility: such as Si, and K, leading to higher proportion of quartz-normative basalts on the one hand and alkalinity on the other. However, the Soviet Mars 5 gamma-ray experiment indicates a K/U lower than terrestrial by one-third (Anders and Owen, 1977). An important volatile effect on Mars discussed in section 10.3.3 is the weathering of martian volcanics, complicating interpretation of remote-sensing data.

The differentiated asteroids should afford the best opportunity to view pristine primordial basalts and their contexts. If the chondritic meteorites are any guide, the range of oxidation level and volatile content among asteroids should be wide.

As discussed in section 10.3.6, Europa and Io manifest the effects of high volatile content, resulting in volcanic activity not easily compared to basaltic volcanism.

10.4.4 Outstanding problems and directions for the future

The obvious way to answer many of the questions about basaltic volcanism on the other planets is to go there and look, meanwhile having improved techniques on how to look. However, other questions require deeper understanding to answer. Significantly, most of these questions pertain to basaltic volcanism on the Earth itself. The following list represents a consensus of the authors of this volume as to these problems and, in some cases, their possible solutions.

(1) What is the relative quantitative significance of different basalt types on Earth? How do they vary with tectonic context, upper mantle composition, and other factors? The suites assembled and described in Chapter 1 are still a sampling, and much work remains to be done to achieve a comprehensive quantitative mapping of basalts by composition, volume, and epoch. The diversity as well as the accessibility of the Earth makes it the fundamental basis for understanding igneous differentiations on other planets.

(2) What is the evolution of igneous rocks with time and how does it constrain the Earth's evolution? So far, the only distinct change in composition determined is the shift from Archean to post-Archean some 2.5 b.y. ago. Surely there should be subtler trends inferrable from suites generated at intermediate epochs.

(3) How can the different effects on basaltic composition—source composition, percent melting, crystallization, mixing, fluid separation—be better discriminated? Most of the improvements needed are fairly evident: phase equilibria experiments at higher pres-

sures; experiments on five or more components, and better means to represent the results; better trace element partitioning coefficients, and their application to samples; better physical models of melting, magma fractionation, etc. The problem seems more to decide among these many possibilities.

(4) How can basaltic volcanism on Earth, and its relationship to other (intermediate and acidic) volcanism be used to predict conditions on other planets? Are some of the less abundant types more sensitive indicators of differences in volatile content, temperature, etc.?

(5) How are isotopic ratios affected by the various processes affecting basalts, including post-eruptive alterations as well as magmatic differentiations? The significance of isotopic measurements—the events discriminated, etc.—is often uncertain, because of complex local circumstances: contamination, stability of mineral sites, metamorphic effects, etc.

(6) What is the relationship of convection to magmatism, and of magmatism to volcanism? There is a crude positive correlation of volcanism with indicators of convective upstreams, such as positive gravity anomalies. However, within a broad "hot spot" area, volcanism seems to shift sporadically from one site to another, and from one composition to another. Obviously, local heterogeneities in composition and temperature play some role, but how these interact with changing physical circumstances [see (7) below] is not clear. Also quite uncertain is the portion of each magmatic event which reaches the surface.

(7) How does magma penetrate the lithosphere? To what extent is it dependent on larger scale tectonic setting? Various aspects of the physics of magma ascent have been modelled, but all such models were severely hypothesis-dependent. Observation, as well as thermal considerations, suggest that volatiles may be significant in enabling a magma to penetrate to the surface. The volumetric domination of ocean rise basalts on the Earth, the predominantly tensional regime of Mars, and the shutting off of volcanism on the Moon and Mercury (since entering their compressive phases) make it evident that macroscale tectonics are important.

(8) How can isotopic indicators of mantle inhomogeneity be improved, and how can they be used to constrain convective and evolutionary models? Several problems are entangled here, such as the refinement of the temporal restrictions of some isotopic systems, the definition of the volumes represented by isotopic samples, the implication of macroscale fluid dynamical modellings for local elements of matter, and differing migration rates of solid and gaseous components. Most of these difficulties will probably be solved in a one-by-one way, but a couple of decades hence the combination of isotopic measurements with thermal convective modellings should give a much better picture of mantle evolution: the degree of interchange between reservoirs, the relationship of heat transfer to material transfer, the principal differentiations, the rate of outgassing, etc.

(9) What are the physical properties of the deeper mantle, and the resulting thermomechanical state? Not only are the effects of phase transitions on viscosity, thermal expansion, and other properties below ~400 km imperfectly known or unknown, the implications of these changes in physical properties are not well understood; as, for example, the relationship of temperature dependence of viscosity to the convective tendency to drive the temperature curve to an adiabat.

(10) How can the interplanetary correlation of volcanic chronologies be improved? The relationship between the cratering rates on different planets is an analytic problem which is model-dependent: it appears to require definition of the reservoirs in which small bodies have been kept, and identification of the mechanism(s) by which these bodies are transferred to trajectories which cross planetary orbits. Possibly, ways will be found to short-circuit some of these dynamical difficulties to give a more effective interplanetary chronology.

(11) What are the thermodynamics and hydrodynamics of great impacts? Better understanding of these processes is important to several problems, such as early heating of the planets and planetesimals, and thence their differentiation; and the evolution of the surfaces of the Moon, Mercury and Mars. Fairly simple considerations indicate that the scaling up from small impacts currently feasible for experiment or computation is by no means simple, and that the energy partitionings, fragment sizes, and material effects may change appreciably.

ACKNOWLEDGMENTS—We are grateful to R. J. Arculus, P. M. Cassen, W. K. Hartmann, R. J. Phillips, S. C. Solomon, D. M. Walker, G. W. Wetherill, D. E. Wilhelms, and J. A. Wood for detailed reviews. The following support for work on Chapter 10 was received: W. M. Kaula, NASA grant NGL 05-007-002; D. L. Anderson, NASA grants NSG-7610 and NGL 05-007-069.

10.5 REFERENCES

Anders E. (1976) Chemical compositions of the Moon, Earth and eucrite parent body. *Phil. Trans. Roy. Soc. Lond.* **A285**, 23–40.

Anders E. and Owen T. (1977) Mars and Earth: origin and abundance of volatiles. *Science* **198**, 453–465.

Anderson D. L. (1975) Chemical plumes in the mantle. *Geol. Soc. Amer. Bull.* **86**, 1593–1600.

Anderson D. L. (1977) Composition of the mantle and core. *Ann. Rev. Earth Planet. Sci.* **5**, 179-202.

Anderson D. L. (1979) The upper mantle transition region: eclogite? *Geophys. Res. Lett.* **6**, 433-435.

Birch F. (1972) Elasticity and constitution of the Earth's interior. *J. Geophys. Res.* **52**, 277-286.

Butler R. F. (1972) Natural remanent magnetization and thermomagnetic properties of the Allende meteorite. *Earth Planet. Sci. Lett.* **17**, 120-128.

Cameron A. G. W. (1978a) Physics of the primitive solar disk. *Moon and Planets* **18**, 5-40.

Cameron A. G. W. (1978b) The primitive solar accretion disk and the formation of the planets. In *The Origin of the Solar System* (S. F. Dermott, ed.), p. 49-73. John Wiley, N.Y.

Cameron A. G. W. and Truran J. W. (1977) The supernova trigger for formation of the solar system. *Icarus* **30**, 447-461.

Cameron A. G. W. and Ward W. R. (1976) The origin of the Moon (abstract). *Lunar Science VII*, p. 120-122. The Lunar Science Institute, Houston.

Campbell D. B., Dyce D. B., Ingalls R. P., Pettengill G. H. and Shapiro I. I. (1972) Venus: topography revealed by radar data. *Science* **175**, 514-516.

Chapman C. R. (1979) The asteroids: nature, interrelations, origin, and evolution. In *Asteroids* (T. Gehrels, ed.), p. 25-60. Univ. Arizona Press, Tucson.

Clayton R. N. and Mayeda T. K. (1975) Genetic relations between the Moon and meteorites. *Proc. Lunar Sci. Conf. 6th*, p. 1761-1769.

Clayton R. N., Onuma N. and Mayeda T. K. (1976) A classification of meteorites based on oxygen isotopes. *Earth Planet. Sci. Lett.* **30**, 10-18.

Consolmagno G. J. and Drake M. J. (1977) Composition and evolution of the eucrite parent body: evidence from rare-earth elements. *Geochim. Cosmochim. Acta* **41**, 1271-1282.

Dickinson W. R. and Luth W. C. (1971) A model for plate tectonic evolution of mantle layers. *Science* **174**, 400-404.

Drake M. J. (1980) Trace elements as quantitative probes of differentiation processes in planetary interiors. *Rev. Geophys. Space Phys.* **18**, 11-26.

Elmegreen B. C. (1978) On the interaction between a strong stellar wind and a surrounding disk nebula. *Moon and Planets* **19**, 261-278.

Fish R. A., Goles G. G. and Anders E. (1960) The record in the meteorites. III On the development of meteorites in asteroidal bodies. *Astrophys. J.* **132**, 243-258.

Frey F. A., Green D. H. and Roy S. D. (1978) Integrated models of basalt petrogenesis: a study of quartz tholeiites to olivine melilites from S.E. Australia utilizing geochemical and experimental petrological data. *J. Petrol.* **19**, 463-573.

Goldreich P. and Ward W. R. (1973) The formation of planetesimals. *Astrophys. J.* **183**, 1051-1061.

Goldstein J. R. and J. M. Short (1967) The iron meteorites, their thermal history and parent bodies. *Geochim. Cosmochim. Acta* **31**, 1733-1770.

Gray C. M. and Compston W. (1976) Excess ^{26}Mg in the Allende meteorite. *Nature* **251**, 495-497.

Green D. H., Hibberson W. O. and Jaques A. L. (1979) Petrogenesis of mid-ocean ridge basalts. In *The Earth: its Origin, Structure and Evolution* (M. W. McElhinny, ed.), p. 265-299. Academic Press, N.Y.

Greenberg R., Wacker J. F., Hartmann W. K. and Chapman C. R. (1978) Planetesimals to planets: numerical simulation of collisional evolution. *Icarus* **35**, 1-26.

Hager B. H. and O'Connell R. J. (1979) Kinematic models of large-scale flow in the Earth's mantle. *J. Geophys. Res.* **84**, 1031-1048.

Harris A. W. (1978) Dynamics of planetesimal formation and planetary accretion. In *The Origin of the Solar System* (S. F. Dermott, ed.), p. 469-492. John Wiley, N.Y.

Hartmann W. K. and Davis D. (1975) Satellite-sized planetesimals and lunar origin. *Icarus* **24**, 504-515.

Hedge C. F. (1978) Strontium isotopes in basalts from the Pacific Ocean basin. *Earth. Planet. Sci. Lett.* **38**, 88-94.

Hoffman J. H., Hodges R. R., McElroy M. B., Donahue T. M. and Kolpin M. A. (1979) Venus lower atmospheric composition: preliminary results from Pioneer Venus. *Science* **203**, 800-802.

Hörz F. (1978) How thick are mare basalts? *Proc. Lunar Sci. Conf. 9th*, p. 3311-3331.

Kaula W. M. (1975) The seven ages of a planet. *Icarus* **26**, 1-15.

Kaula W. M. (1977) On the origin of the Moon, with emphasis on bulk composition. *Proc. Lunar Sci. Conf. 8th*, p. 321-331.

Kaula W. M. (1979) Thermal evolution of Earth and Moon growing by planetesimal impacts. *J. Geophys. Res.* **84**, 999-1008.

Kaula W. M. (1980a) The beginning of Earth's thermal evolution. In *The Continental Crust and its Mineral Deposits* (D. Strangway, ed.), Geol. Assoc. Canada, Spec. Paper 20, p. 25-34.

Kaula W. M. (1980b) Material properties for mantle convection consistent with observed surface fields. *J. Geophys. Res.* **85**, 7031-7044.

Kaula W. M. and Bigeleisen P. E. (1975) Early scattering by Jupiter and its collision effects in the terrestrial zone. *Icarus* **25**, 18-33.

Kaula W. M. and Harris A. W. (1975) Dynamics of lunar origin and orbital evolution. *Rev. Geophys. Space Phys.* **13**, 363-371.

Kieffer S. W. and Simonds C. H. (1980) The role of volatiles and lithology in the impact cratering process. *Rev. Geophys. Space Phys.* **18**, 143-181.

Lee T., Papanastassiou D. A. and Wasserburg G. J. (1976) Demonstration of ^{26}Mg excess in Allende and evidence for ^{26}Al. *Geophys. Res. Lett.* **3**, 109-112.

Liebermann R. C. and Ringwood A. E. (1976) Elastic properties of anorthite and the nature of the lunar crust. *Earth. Planet. Sci. Lett.* **31**, 69-74.

Lynden-Bell D. and Pringle J. E. (1974) The evolution of viscous discs and the origin of the nebular variables. *Mon. Not. Roy. Astron. Soc.* **168**, 603-637.

McGetchin T. R. and Smyth J. R. (1978) The mantle of Mars: some possible geological implications of its high density. *Icarus* **34**, 512-536.

Morgan J. W., Higuchi H., Takakashi H. and Hertogen J. (1978) A "chondritic" eucrite parent body: inference from trace elements. *Geochim. Cosmochim. Acta* **42**, 27-38.

O'Hara M. J. (1968) Are ocean-floor basalts primary magmas? *Nature* **220**, 683-686.

O'Keefe J. A. and Sullivan F. (1978) Fission origin of the Moon: cause and timing. *Icarus* **35**, 272-283.

O'Keefe J. D. and Ahrens T. J. (1977) Impact-induced energy partitioning, melting, and vaporization on terrestrial planets. *Proc. Lunar Sci. Conf. 8th*, p. 3357-3374

O'Nions R. K., Evensen N. M. and Hamilton P. J. (1979) Geochemical modeling of mantle differentiation and crustal growth. *J. Geophy. Res.* **84**, 6091-6101.

Peale S. J., Cassen P. and Reynolds R. T. (1979) Melting of Io by tidal dissipation. *Science* **203**, 892-894.

Pettengill G. H., Eliason E., Ford P. G., Loriot G. B., Masursky H. and McGill G. E. (1980) Pioneer Venus radar results: altimetry and surface properties. *J. Geophys. Res.* **85**, 8261-8270.

Pettengill G. H., Ford P. G., Brown W. E., Kaula W. M., Keller C. H., Masursky H. and McGill G. E. (1979a) Pioneer Venus Radar mapper experiment. *Science* **203**, 806-808.

Pettengill G. H., Ford P. G., Brown W. E., Kaula W. M., Keller C. H., Masursky H. and McGill G. E. (1979b) Venus: preliminary topographic and surface imaging results from the Pioneer Orbiter. *Science* **205**, 90-93.

Phillips R. J., Kaula W. M., McGill G. E. and Malin M. C. (1981) Tectonics and evolution of Venus. *Science*, in press.

Pollack J. B. (1979) Climatic change on the terrestrial planets. *Icarus* **37**, 479-553.

Ringwood A. E. (1975) *Composition and Petrology of the Earth's Mantle.* McGraw-Hill, N.Y. 618 pp.

Ringwood A. E. (1979) *Origin of the Earth and Moon.* Springer-Verlag, Berlin. 295 pp.

Ruskol E. L. (1977) The origin of the Moon. In *The Soviet-American Conference on Cosmochemistry of the Moon and Planets*, p. 815-822. NASA, SP-370. Washington, D.C.

Safronov V. S. (1972) *Evolution of the Protoplanetary Cloud and Formation of the Earth and Planets.* NASA TTF-667. Washington, D.C. 206 pp.

Schaber G. G. (1973) Lava flows from Mare Imbrium: geologic evolution from Apollo orbital photography. *Proc. Lunar Sci. Conf. 4th*, p. 73-92.

Schilling J. G. (1971) Sea-floor evolution: rare-earth evidence. *Phil. Trans. Roy. Soc. Lond.* **A286**, 663-706.

Schramm D. N. and Wasserburg G. J. (1970) Nucleochronologies and the mean age of the elements. *Astrophys. J.* **162**, 57-69.

Sjogren W. L., Phillips R. J., Birkeland P. W. and Wimberly R. N. (1980) Gravity anomalies on Venus. *J. Geophys. Res.* **85**, 8295-8302.

Slattery W. L. (1978) Protoplanetary Core formation by rainout of iron drops. *Moon and Planets* **19**, 443–456.

Slattery W. L., DeCampli W. M. and Cameron A. G. W. (1980) Protoplanetary core formation by rain-out of minerals. *Harvard Cen. Astrophys. Preprint 1280.* 9 pp.

Smith J. V. (1979) Mineralogy of the planets: a voyage in space and time. *Mineral. Mag.* **43**, 1–89.

Solomon S. C. and Chaiken J. (1976) Thermal expansion and thermal stress in the Moon and terrestrial planets: clues to early thermal history. *Proc. Lunar Sci. Conf. 7th*, p. 3229–3243.

Solomon S. C. and Head J. W. (1979) Vertical movement in mare basins: relation to mare emplacement, basin tectonics, and lunar thermal history. *J. Geophys. Res.* **84**, 1667–1682.

Sonett C. P. and Herbert F. L. (1977) Pre-main sequence heating of planetoids. In *Comets, Asteroids, and Meteorites* (A. Delsemme, ed.), p. 429–437. Univ. of Toledo Press.

Stevenson D. J. (1980) Lunar asymmetry and paleomagnetism. *Nature* **287**, 520–521.

Stevenson D. J. and Turner J. S. (1979) Fluid models of mantle convection. In *The Earth: Origin, Evolution, and Structure* (M. McElhinny, ed.), p. 227–263. John Wiley, N.Y.

Stevenson D. J., Schubert G., Cassen P. and Reynolds R. J. (1980) Core evolution and magnetism of the terrestrial planets (abstract). In *Lunar and Planetary Science XI*, p. 1091–1093. Lunar and Planetary Institute, Houston.

Stolper E. (1977) Experimental petrology of eucritic meteorites. *Geochim. Cosmochim. Acta* **41**, 587–611.

Stolper E. (1980) Predictions of mineral assemblages in planetary interiors. *Proc. Lunar Sci. Conf. 11th*, p. 235–250.

Strom S. E. (1977) Early stages of stellar evolution. In *Star Formation* (T. DeJong and A. Maeder, eds.), p. 179–197. Reidel, Dordrecht.

Sun S. -S. and Hanson G. N. (1975) Evolution of the mantle: geochemical evidence from alkali basalts. *Geology* 3, 297–302.

Tatsumoto M. (1978) Isotopic composition of lead in oceanic basalt and its implication to mantle evolution. *Earth Planet. Sci. Lett.* **38**, 63–87.

Trimble V. (1975) The origin and abundances of the chemical elements. *Rev. Mod. Phys.* **47**, 877–976.

Ulrich R. K. (1978) The status of T. Tauri models. In *Protostars and Planets* (T. Gehrels, ed.), p. 718–773. Univ. of Arizona Press, Tucson.

Urey H. C. (1956) Diamonds, meteorites, and the origin of the solar system. *Astrophys. J.* **124**, 623–637.

Vinogradov A. P., Surkov Y. A. and Kimozov F. F. (1973) The contents of uranium, thorium, and potassium in the rocks of Venus as measured by Venera 8. *Icarus* **20**, 253–259.

Walker D., Stolper E. M. and Hays J. F. (1979) Basaltic volcanism: the importance of planet size. *Proc. Lunar Sci. Conf. 10th*, p. 1995–2015.

Walker J. C. G. (1975) Evolution of the atmosphere of Venus. *J. Atmos. Sci.* **32**, 1248–1256.

Warren P. H. and Wasson J. T. (1979) The origin of KREEP. *Rev. Geophys. Space Phys.* **17**, 73–88.

Wasserburg G. J. and DePaolo D. J. (1979) Models of earth structure inferred from neodymium and strontium isotopic abundances. *Proc. Natl. Acad. Sci. USA* **76**, 3594–3598.

Wasson J. T. and Warren P. H. (1980) Contribution of the mantle to the lunar asymmetry. *Icarus*, in press.

Wasson J. T. and Wetherill G. W. (1979) Dynamical, chemical and isotopic evidence regarding the formation locations of asteroids and meteorites. In *Asteroids* (T. Gehrels, ed.), p. 926–974. Univ. of Arizona Press, Tucson.

Weidenschilling S. J. (1975) Mass loss from the region of Mars and the asteroid belt. *Icarus* **26**, 361–366.

Weidenschilling S. J. (1978) Iron/silicate fractionation and the origin of Mercury. *Icarus* **35**, 99–111.

Wetherill G. W. (1976) The role of large bodies in the formation of the Earth and Moon. *Proc. Lunar Sci. Conf. 7th*, p. 3245–3257.

Wetherill G. W. (1980a) Numerical calculations relevant to the accumulation of the terrestrial planets. In *The Continental Crust and its Mineral Deposits*. (D. Strangway, ed.), Geol. Assoc. Canada, Spec. Paper 20, in press.

Wetherill G. W. (1980b) Could the solar wind have been the source of the high concentration of ^{36}Ar in the atmosphere of Venus? (abstract). In *Lunar and Planetary Science XII*, p. 1239–1241. Lunar and Planetary Institute, Houston.

Wood J. A. (1964) The cooling rates and parent planets of several iron meteorites. *Icarus* 3, 429–459.